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FEBRUARY 1959

PLEISTOCENE GLACIERS ON SOUTHERN CALIFORNIA MOUNTAINS*

ROBERT P. SHARP[1], CLARENCE R. ALLEN[1], and MARK F. MEIER[2]

ABSTRACT. The deposits of 7 valley glaciers have been mapped in the San Gorgonio area of the San Bernardino Mountains in southern California. These ice bodies headed at elevations between 10,300 and 11,300 feet, the lowest elevation attained was 8700 feet, and lengths were 0.5 to 1.7 miles. Dry Lake glacier was the largest. It covered 0.84 square miles on the north slope of San Gorgonio Mountain.

The principal products of glaciation are cirques and huge terminal embankments of coarse angular debris up to 700 feet high. Sharp-crested end moraines, typical recessional loops, and a number of other relations indicate that these embankments were formed by glaciers rather than as rock glaciers or debris flows. Two separate episodes of glaciation are recognized; both are considered Wisconsin.

Features previously attributed to glaciers in the San Gabriel Mountains were restudied. It is concluded that this range escaped glaciation in the Wisconsin and probably all earlier Pleistocene stages. San Jacinto Peak may have been glaciated, but definite proof has not yet been found.

INTRODUCTION

Evidence of glaciation on the slopes of San Gorgonio Mountain[3] has previously been reported (Fairbanks and Carey, 1910; Vaughan, 1922, p. 335-336). New maps, air photos, developments in the knowledge of mountain glaciation, and international interest in this occurrence (Klebelsberg, 1949, p. 463, 510; Flint, 1957, p. 310) motivated restudy of this and other possible sites of glaciation on southern California mountains. Glaciers in marginal areas are particularly sensitive indicators of environmental change and afford a possible means of establishing links with other Pleistocene events. Field work amounting to 29 man-days was carried on principally during weekends from May of 1954 to December of 1957.

PHYSICAL SETTING

For this report, southern California is arbitrarily defined as that part of the state south of Latitude 35° N. This eliminates high peaks of the southern Sierra Nevada and Telescope Peak in the Panamint Range. San Gorgonio Mountain (11,502 feet), the highest point in southern California, is in the southeastern part (34° 06′ N., 116° 49′ W.) of the San Bernardino Mountains (fig. 1). To the southeast across San Gorgonio Pass is San Jacinto Peak (10,831 feet), and to the west is the San Gabriel Range, which attains an elevation of 10,080 feet.

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[3] Reprint of a paper on this subject by Ingle and Moran (1958) was received after the present article had been set in type.

Fig. 1. Map of southern California localities.

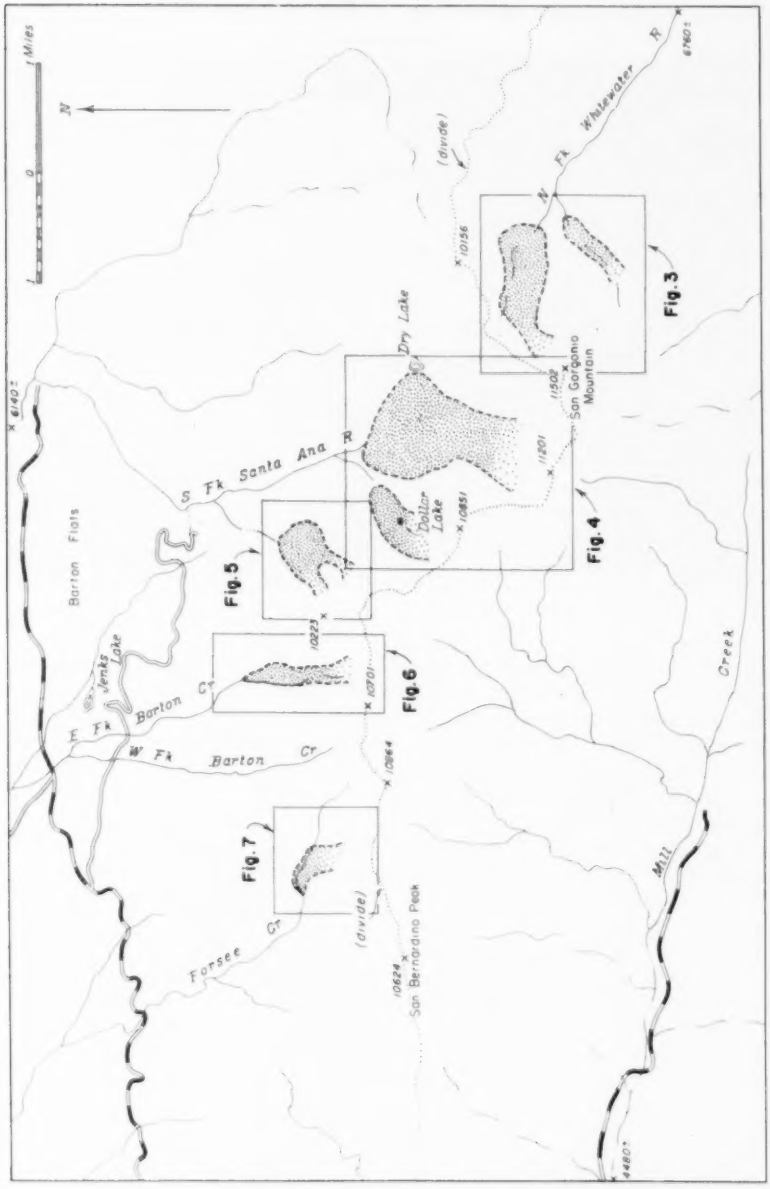


Fig. 2. Glaciated sites in the San Bernardino Mountains.

GLACIATION OF THE SAN GORGONIO MOUNTAIN AREA

Introduction

Localities of principal interest are on San Gorgonio Mountain and along a high ridge extending 6 miles west to San Bernardino Peak, shown on the new San Gorgonio Mountain quadrangle (1/62,500, 1955). Most elevations along this ridge are between 10,400 and 11,200 feet. The average annual precipitation above 8500 feet on San Gorgonio Mountain probably approaches 50 inches (Troxell, 1954, plate 6A), much of which is snow. Under present conditions snow banks linger well into July in protected places high on north-facing slopes.

The Glaciers

At least 7 cirque and valley glaciers formed in the San Gorgonio Mountain area (fig. 2). They headed at elevations between 10,300 and 11,300 feet, and none descended below 8700 feet (table 1). Lengths were 0.5 to 1.7 miles and areas 0.1 to 0.84 square miles. The Dry Lake glacier on the north side of San Gorgonio Mountain was the largest. Steep gradients, as much as 800 to 1000 feet per mile, prevented the ice from attaining a thickness in excess of a few hundred feet.

Consideration has been given to the possibility that the deposits and as-

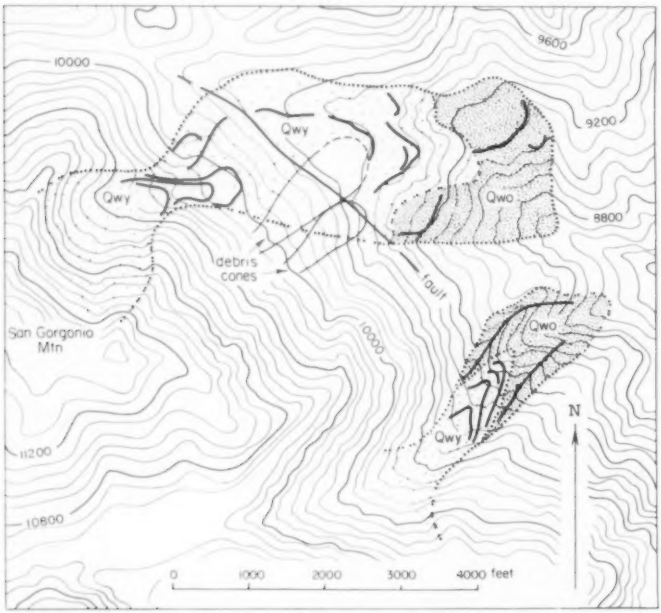


Fig. 3. Glacial deposits in the North Fork of Whitewater River area. South Prong at lower right. Qwo = old episode and Qwy = young episode of glaciation. Heavy lines represent individual moraines. Same designations apply for figures 3 to 7 inclusive.

PLATE 1
Moraines

Limbs of two end-moraine loops of the younger episode on east side of South Prong of North Fork of Whitewater River.

sociated topographic features might have formed as rock glaciers or debris flows. They are judged to be glacial for the following reasons: (1) The younger end moraines are distinct sharp-crested ridges and not rounded wrinkles or the ends of a series of bulbous concentric lobes such as characterize rock glaciers and debris-flow tongues. (2) Many of the end moraines are colinear with and clearly related to lateral moraines; this is generally not true of the swells or wrinkles on rock glaciers or debris-flow tongues. (3) The deposits inside the lateral ridges of a debris flow in general have a transverse profile convex upward. The deposits described herein have concave upward profiles. (4) Successive debris flows piled one on top of the other form embanked ridges along the centers of which the latest flows have run. There are no such features here. (5) The relatively narrow range in the upper and lower elevation limits for these features and their localization on northward-facing slopes make landslides an unlikely explanation. (6) The deposits are too blocky and lacking in fines to be typical debris-flow products.

The relations at all the glaciated sites suggest two separate glacial episodes of significantly different age. From east to west the glaciers occupied cirques or canyon heads as follows (fig. 2):

South Prong of North Fork of Whitewater River.—At the mouth of this canyon (fig. 3) are two prominent ridges which morphologically and constitutionally have the characteristics of lateral moraines. Plastered on the inside of the eastern lateral ridge is a distinctly younger, sharp-crested ridge, and

nestled inside and near the head of these ridges are at least 4 and possibly 6 end moraines (plate 1) of the younger glacial episode (fig. 3). Much of the nourishment for this glacier probably gathered in a cirque on the north face of a peak nearly 11,000 feet high on the ridge south of this canyon (plate 2A). Some snow was also supplied from the steep south face of San Gorgonio Mountain.

North Fork of Whitewater River.—This glacier accumulated in a large cirque (plate 2A) on the northeast face of San Gorgonio Mountain (fig. 3). Its terminal position is marked by a bulky embankment completely choking the canyon and rising fully 700 feet above its east base, where several springs emerge. On the tread above this riser is an irregular complex of knobs, short ridges and closed depressions, cut transversely by a recent fault scarplet (fig. 3). A large lateral moraine projecting from the northeast ridge of the cirque encloses a number of loop moraines of the younger episode. The cirque is choked with coarse talus, but near its mouth are two distinct recessional moraines.

Dry Lake.—The glacial deposits of this locality have the greatest bulk, area, and complexity (fig. 4). The bulbous glacier which formed them was fed by two ice streams draining from large cirques on the north face of San Gogonio Mountain (plate 2B), or more specifically Jepson Peak (fig. 4). The huge embankment of coarse debris left by this ice closes the basin of Dry Lake on its west side and rises steeply 650 feet above South Fork Meadows. Many springs emerge from the base of this porous material. The topography of the tread at the top of this embankment is typically glacial, featuring knobs, depressions, and long curved morainal ridges with a relief of 25 to 100 feet. Flattish areas of outwash are scattered between the prominences. Well-shaped lateral moraines define the west margin, and high in the cirques are some recessional loops. Topographic complexity may be due partly to the fact that the two ice streams which formed the Dry Lake glacier bulb maintained a degree of independent behavior and produced a complex system of narrow loops which are much generalized by the scale of figure 4.

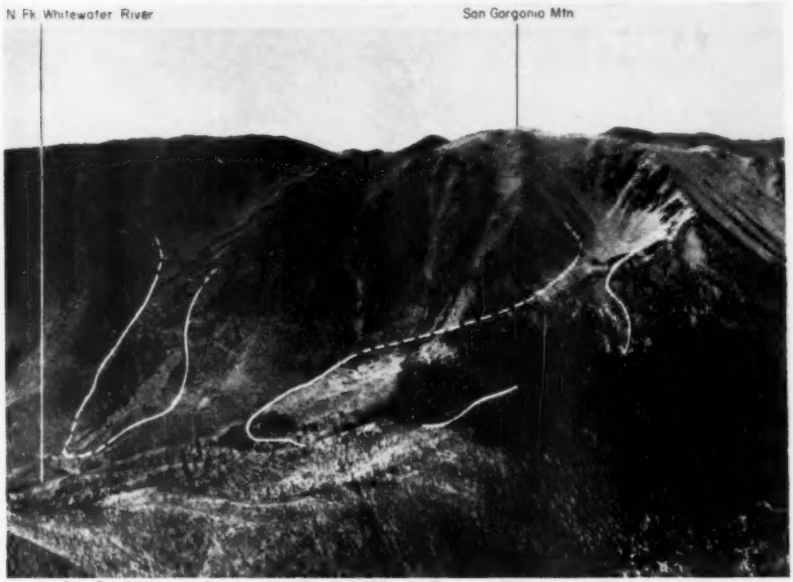
Dollar Lake.—The canyon of Dollar Lake is choked by a large embankment of coarse debris, and the lake is enclosed by a strong end moraine continuous with a lateral moraine extending 0.5 mile to the west on the north side of the canyon (fig. 4). Closure in the lake basin approaches 100 feet. The contact between deposits of the older and younger glaciations is marked by a small step part way down the face of the morainal embankment. Morainal loops and lateral ridges are easily recognized in the older deposits, but the loops are breached by stream erosion. Much of the nourishment for this glacier came from the north face of Charlton Peak.

West Prong of South Fork of Santa Ana River.—This valley held two small cirque glaciers of steep gradient which merged to build a single large embankment of debris during the older glaciation (fig. 5).

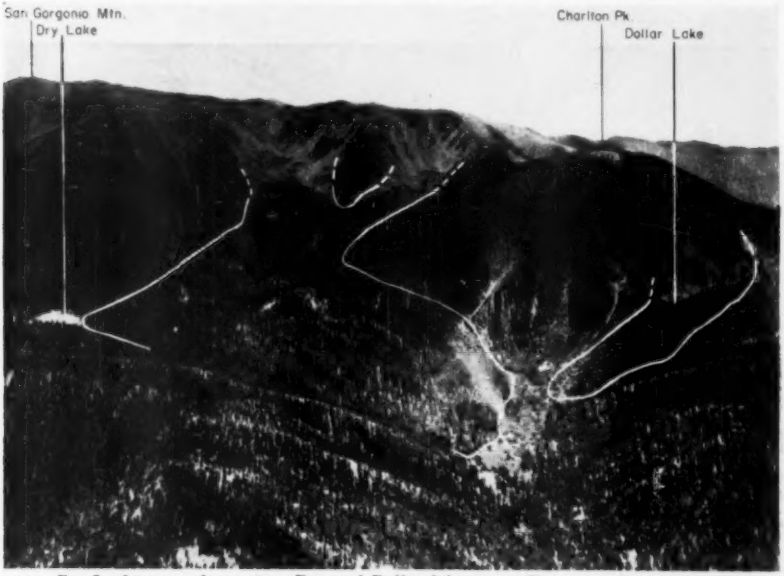
East Fork of Barton Creek.—The glacial deposits in this canyon are among the most instructive of the area. They feature a 700-foot high bouldery embankment choking the canyon from wall to wall (fig. 6). The north face of this embankment, consisting of material laid down during the older episode,

PLATE 2

Air Photos of San Gorgonio Mountain Glaciated Sites



A. Looking southwest into head of North Fork of Whitewater River. Cirque at left fed glacier in South Prong and cirque at right fed glacier in North Fork of Whitewater River. Lateral moraines of older stage visible at mouth of South Prong. Irregular topography of burned area in left foreground underlain partly by deposits of North Fork glacier.



B. Looking southwest into Dry and Dollar lakes area. Two central cirques on north face of San Gorgonio Mountain fed Dry Lake glacier. Note complex mass of morainal debris west of Dry Lake. Dollar Lake glacier fed from steep north face of Charlton Peak at right.

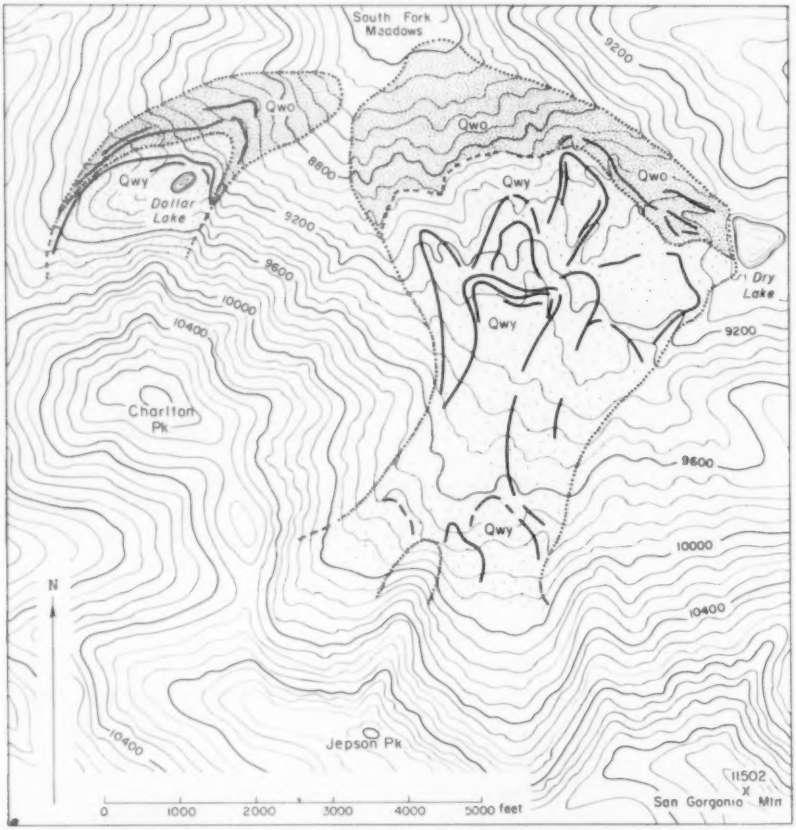


Fig. 4. Glacial deposits in the Dry and Dollar lakes area.

is terminated by a distinct but breached end moraine with springs at its base. The uppermost part of the embankment consists of younger deposits including an exceedingly blocky terminal moraine enclosing a marked axial depression that extends upvalley a quarter of a mile to a recessional moraine. Deposits of the two glaciations are clearly distinguished here by physical appearance and vegetative cover.

Viewed from the air the accumulations on Barton Creek are suggestive of rock glaciers. That the deposits are truly glacial is indicated by: (1) the elongate axial depression enclosed by the terminal and its associated lateral moraines of the younger episode, (2) the well-defined recessional moraine higher up the valley, and (3) the long east-side lateral moraine of the older glaciation.

TABLE 1

Data on Pleistocene Glaciers of the San Bernardino Mountains

Locality	Estimated elevation at head of glacier in feet	Lowest elevation attained	Length in miles	Approximate area square miles*	Principal direction of exposure
South Prong, North Fork of Whitewater River	11,000	8700 (1) 9100 (2)	1.0 (1) 0.7 (2)	0.17 (1) 0.14 (2)	E and N
North Fork of Whitewater River	11,300	9120 (1) 9440 (2)	1.3 (1) 1.0 (2)	0.36 (1) 0.30 (2)	N and NE
Dry Lake area	10,900	8800 (1) 9000 (2)	1.7 (1) 1.6 (2)	0.84 (1) 0.74 (2)	N
Dollar Lake	10,400	8960 (1) 9240 (2)	0.7 (1) 0.5 (2)	0.16 (1) 0.12 (2)	N and E
Tributary of South Fork Santa Ana River	10,300	8960 (1) 9000 (2)	0.8 (1) 0.7 (2)	0.19 (1) 0.16 (2)	NE
East Fork Barton Creek	10,300	8800 (1) 9050 (2)	0.8 (1) 0.6 (2)	0.11 (1) 0.09 (2)	N
Forsee Creek	10,300	8960 (1) 9060 (2)	0.6 (1) 0.5 (2)	0.12 (1) 0.10 (2)	NW

(1) = older episode
(2) = younger episode

* Second figure beyond decimal point is reported to indicate approximate difference in size.

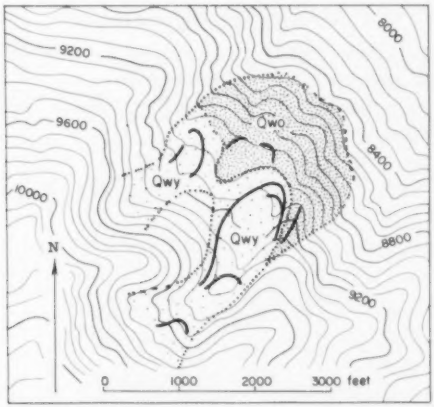


Fig. 5. Glacial deposits in the West Prong of the South Fork of Santa Ana River.

Forsee Creek.—The ice in Forsee Creek lay largely on the south side of the valley under a steep north-facing wall and flowed more across the valley than down it (fig. 7). This glacier built 3 distinct morainal ridges on the north side of the valley and a terminal embankment a short distance downstream. The north-side ridges are too far from the south valley wall to be protalus ramparts.

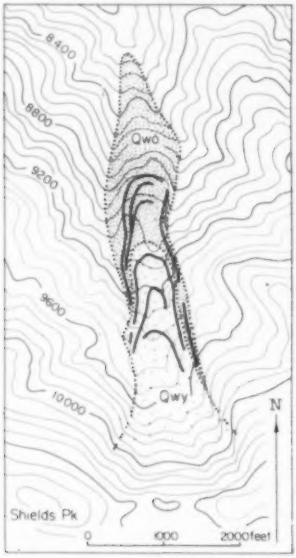


Fig. 6. Glacial deposits in the East Fork of Barton Creek.

Criteria of Age Differentiation

The older and younger glacial deposits of the San Gorgonio Mountain region are differentiated by the following criteria:

1. *Topographic position.*—Older deposits are farther down the valley and higher on the valley walls.

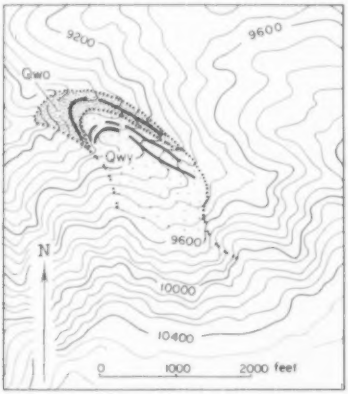


Fig. 7. Glacial deposits in Forsee Creek.

2. *Topographic expression.*—Landforms on the older deposits are rounded and subdued compared to sharp angular features on the younger deposits.

3. *Topographic unconformity.*—In some localities older morainal ridges disappear beneath the younger moraines with a marked difference in trend.

4. *Weathered color.*—Many of the older deposits are yellowish-brown in contrast to a predominantly grayish cast in the younger materials.

5. *Abundance of gruss.*—Gruss formed by disintegration of granitic boulders is more abundant and finer grained on the older deposits. It fills depressions thus helping to create a more subdued topography.

6. *Boulder relations.*—(a) Boulders on younger moraines are more angular and sharp edged because of recent fracturing without much subsequent rounding by other weathering. (b) On older deposits the surfaces of boulders are more roughened by pitting and disintegration. (c) More large boulders are found on younger moraines (table 2). (d) Boulder frequency is higher on the younger deposits (table 2). (e) Spalls and shattered boulders are more abundant on younger moraines composed of granitic rocks, for disintegration has largely destroyed such fragments on older deposits (table 2).

7. *Vegetation cover.*—Forests are relatively dense on the older deposits compared with scattered trees and heavy brush on younger moraines.

TABLE 2

Boulder Size, Frequency, and Condition on Older and Younger Moraines

Location	Predominant rock type	Age	Number of boulders 3 feet diameter in a strip 10 feet wide and 100 feet long	Number of boulders 1 ft. in diameter in a similar strip	Fractured boulders
Dry Lake	Quartz Monzonite	Younger	20	195	49%
		Older	8	75	39%
North Fork of Whitewater River	Quartz Monzonite	Younger	—	120	—
		Older	—	90	—
East Fork of Barton Creek	Gneiss	Younger	74	196	20%
		Older	17	108	20%
Dollar Lake	Quartz Monzonite and Gneiss	Younger	6	96	18%
		Older	4	48	15%

Episodes of Glaciation

By means of the criteria listed, two episodes of glaciation have been distinguished at all sites in the San Gorgonio area. The impression is gained that the older deposits have an age at least twice and possibly several times as great as the younger.

In cirques at the head of North Fork of Whitewater River, on the north face of San Gorgonio Mountain above Dry Lake, and at the head of East Fork of Barton Creek are blocky morainal loops which look fresher than most moraines of the younger episode. Upon some, grow stunted lodgepole pines which increment borings show to be 300 to 500 years old. These accumulations must be older than the 1850 moraines of the Sierra Nevada (Matthes, 1942, p. 197), but they may be due to an earlier phase of the "little ice age" (Matthes, 1942, p. 212-214; Holmes and Moss, 1955, p. 642-643) or to some other post-Wisconsin glacial episode. Since definite evidence is lacking, they are provisionally treated as recessional phases of the younger glaciation.

Correlation of the San Gorgonio Mountain sequence with glacial stages in other western mountains is purely speculative, but all the San Gorgonio deposits appear young enough to be Wisconsin. To some degree, they resemble Blackwelder's (1931) Tahoe (early Wisconsin) and Tioga (late Wisconsin) glacial stages of the Sierra Nevada. Thick deposits of boulder-rich, poorly sorted debris filling the canyon of Santa Ana River north of the areas of recognized glaciation may possibly be related to pre-Wisconsin glaciations. Since no proof of a glacial origin has been found, these deposits may actually be the product of accelerated mass wasting related to late Pleistocene episodes of refrigeration.

SUPPOSED GLACIATION OF THE SAN GABRIEL MOUNTAINS

Landforms and deposits at several localities in the San Gabriel Mountains (fig. 1) have been attributed to glaciation (Miller, 1926; 1928, p. 226-227). Skepticism concerning this interpretation arises from the fact that many of these glaciers would have formed at elevations as low as 5000 to 6000 feet. Even though they may have been pre-Wisconsin (Miller, 1926, p. 81), this seems much too low in view of the fact that Wisconsin glaciers in the San Gorgonio area required elevations 5000 to 6000 feet higher.

Not all of Miller's localities were visited, but at those inspected the relations do not require glaciation. Bear Canyon (Mt. Lowe quadrangle), a tributary of the Arroyo Seco, was described as a glacial U-shaped valley (Miller, 1926, p. 78-80). Upper Bear Canyon is indeed relatively open and lacks the narrow crookedness and overlapping spurs of lower Bear Canyon. However, this seems to be due principally to differences in bedrock, the upper part being in readily disintegrating Mt. Wilson diorite (Miller, 1934, map) and the lower part in more resistant Lowe granodiorite and San Gabriel Formation. Supposed glacial deposits on the floor of upper Bear Canyon have all the appearances of an alluvial accumulation.

The same explanation applies for similar relations in Big Santa Anita Canyon (Mt. Wilson quandrangle) near Sturtevant's Camp (Miller, 1928, p. 227). The upper open part of this canyon is in homogeneous Wilson diorite that is susceptible to relatively rapid granular disintegration.

The most intriguing locality is the Crystal Lake area at the head of North Fork of San Gabriel River (Crystal Lake quadrangle). Here, the deposits and topography are of a type that one might readily attribute to either glaciation (Miller, 1957, p. 212) or mass movements. However, the topography is, if anything, too chaotic, and it lacks the morainal loops and ridges so character-

istic of glacial accumulations on San Gorgonio Mountain. Furthermore, none of the topographic features show any relations to possible sources of ice. If of glacial origin, these deposits could hardly be older than Wisconsin. The peaks and ridges enclosing this area attain a maximum elevation of 8500 feet, and the exposure is south. In view of altitude relations on San Gorgonio Mountain, this seems much too low for Wisconsin glaciers. These deposits and the peculiar topography are attributed to landslides and rock falls. Features on the north side of Strawberry Peak (Miller, 1928, p. 227) farther west in the San Gabriels are also judged to be of this origin.

The highest peak in the San Gabriel Range is Mt. San Antonio (10,080 feet). Inspection of the north, northeast and northwest faces of this peak disclosed no evidence of glaciation. The elevation is below that at which glaciers formed on San Gorgonio Mountain, and there is no reason to expect more favorable conditions such as heavier precipitation here. South of the peak, massive bouldery deposits that choke parts of San Antonio Canyon have been mapped as landslides (Ehlig, personal communication).

It is concluded that the San Gabriel Mountains escaped glaciation in the Wisconsin stage, and no evidence of glaciation during pre-Wisconsin stages has been seen.

SAN JACINTO PEAK

This relatively isolated peak, at 10,831 feet, is high enough for Wisconsin ice to have formed, given precipitation comparable to the San Gorgonio area and suitable sites for snow accumulation. Much of the north face of San Jacinto Mountain is so precipitous that snow would avalanche to lower slopes and glaciers could not form. Some topographic features of the summit area suggest glaciation, but definite proof of ice work has not yet been found. A tongue of bouldery debris extends northeast into Round Valley from an amphitheater heading in the saddle south of San Jacinto Peak. This deposit terminates as a bulbous mass at about 9800 feet and shows evidence of two episodes of development, but it lacks the morainal loops and closed depressions so typical of glacial deposits on San Gorgonio Mountain. Still the general relations incline one to regard it as more likely of glacial origin than not.

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GLAUCOPHANE-SCHISTS AND ASSOCIATED ROCKS NEAR VALLEY FORD, CALIFORNIA

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ABSTRACT. Large residual masses of Franciscan glaucophane-schist, eclogite, and associated metamorphic rocks rest upon a basement of Franciscan graywackes and metagraywackes. The residual metamorphic blocks are probably relict masses from a former cover. Various stages in the transformation of graywackes and cherts into glaucophane-rocks have been observed in both basement rocks and residual masses. Progressive glaucophanization of residual blocks of eclogite also occurs, the final mineralogical and chemical composition approaching that of glaucophane-rocks derived from graywackes. Chemical changes accompanying the metamorphism of the graywackes involve an addition of Fe, Mg and Ca, together with a decrease in Si and Na. These changes contrast with those which have produced glaucophane-rocks from the eclogites. The mineralogical character of the glaucophane-schists suggests affinities with the greenschist or epidote-amphibolite facies, but the instability of albite and the presence of hydrous minerals with high density indicates certain conditions perculiar to the glaucophane-schist facies; possibly high water-vapor pressure.

INTRODUCTION AND GENERAL RELATIONSHIPS

The rocks to be described occur in the Sebastopol Quadrangle, 7½ miles north, 17° east of Valley Ford, Sonoma County, on the Valley Ford-Freestone highway (Gealey, 1950; Bloxam, 1955). It is a locality well known for excellent exposures of glaucophane-schist, eclogite, and associated rocks of the Franciscan (Jurassic) formation (Switzer, 1950). The area forms a small part of the Coast Range belt of Franciscan rocks, extending south towards San Francisco and northwards into the Healdsburg Quadrangle (fig. 1).

At Valley Ford the rocks can be divided into three major groups: (a) Basement rocks of the Franciscan which are mainly graywackes, but include metagraywackes, glaucophane-schists, and some igneous rocks. The metamorphic rocks are confined to the area east of a fault which brings them into contact with the graywackes.

(b) Residual masses of Franciscan chert, metachert, glaucophane-schist, eclogite, and associated rocks, resting upon the basement graywackes and metagraywackes. Some of the metamorphic rock-types are common to both basement rocks and residual masses.

(c) A cover of soft unconsolidated marine sands and clays of the Merced Formation (Pliocene), which has in places been stripped off to reveal rocks of the previous two categories.

The residual metamorphic rocks are well exposed and rise above the low-lying ground composed of relatively soft graywackes and metagraywackes (plate 1A). The resulting differential erosion has, in most cases, obscured the original contact relationships. Franciscan metamorphic rocks in this type of setting appear to be common in many parts of California (Gealey, 1950; Borg, 1956).

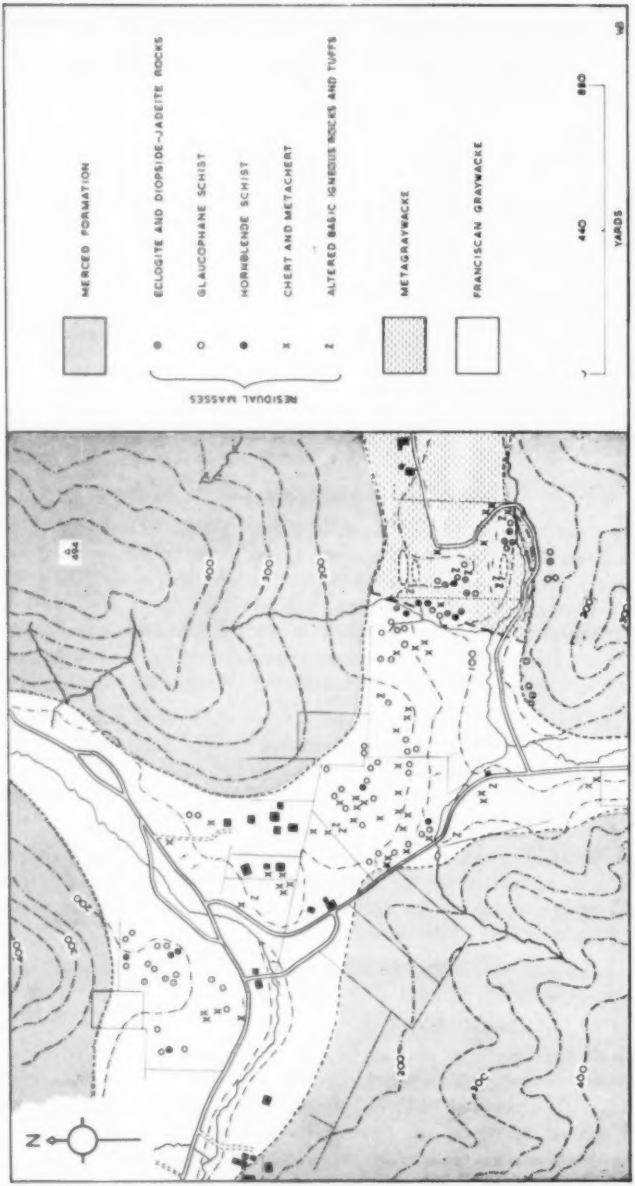
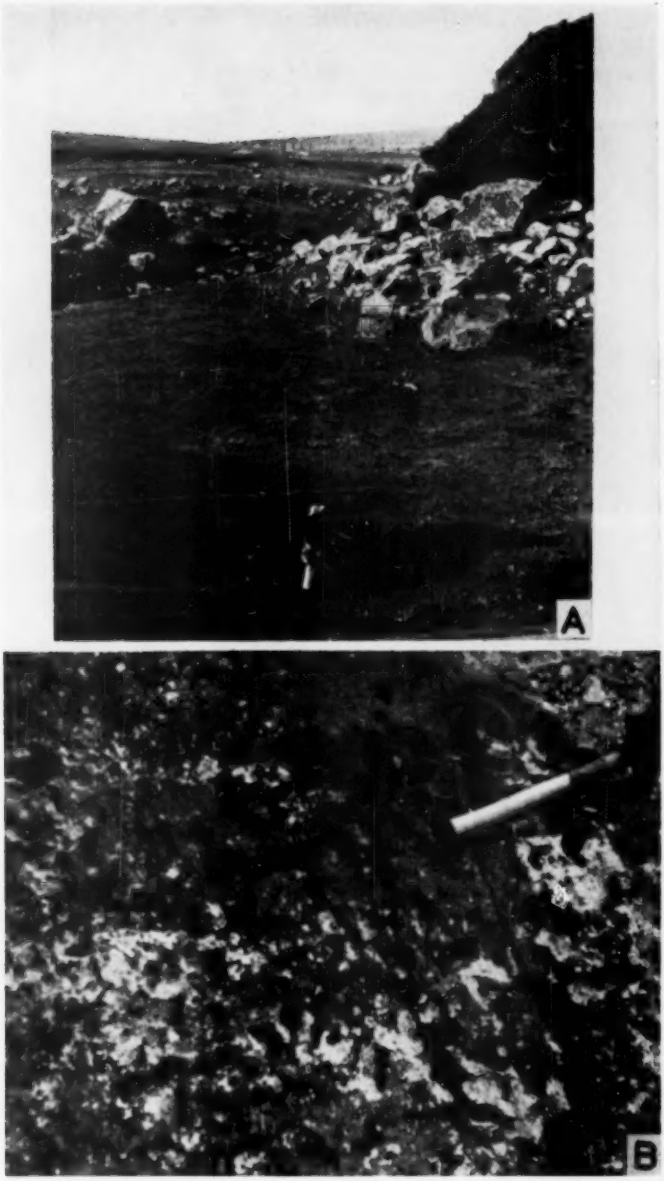


Fig. 1. Geological map of the area near Valley Ford, Sonoma County, California. Major buildings are shown, together with the principal field boundaries as in the fall, 1954.

PLATE 1



A. Residual blocks of Franciscan glaucophane-schist, metachert and associated metamorphic rocks, Valley Ford. The observer is standing on unaltered basement Franciscan graywackes.

B. *Pholas* borings in glaucophane-schist, Valley Ford.

BASEMENT ROCKS

Franciscan graywackes.—In general exposures of the basement rocks are poor and confined to creeks, highway-cuttings, and small quarrys. The dominant rock-type is arkosic graywacke with local pebbly beds. They are strongly folded with fold-axes trending north to north-northwest and plunging at about 10° in the same direction. This structural trend at Valley Ford agrees with the dominant strike of the Franciscan rocks over much of northern California.

The graywackes consist essentially of angular quartz and chert detritus, together with frequent basic igneous fragments, commonly variolitic basalts. Fresh soda-plagioclase, sericite, chlorite, and carbonaceous or shaley material are always present, with soda-plagioclase constituting some 30 percent of the rock in many cases (302/74).*

An analysis of a typical Valley Ford graywacke is given in table 1, 1; where the low value for total Fe, Mg and Ca is indicated. This arkosic tendency appears to be a general feature of the Franciscan graywackes and has been noted by Taliaferro (1942, p. 83) and Bloxam (1956, p. 494).

Some of the apparently unaltered basement graywackes exhibit, under the microscope, signs of incipient metamorphism (302/69). Fragments of variolite and shaley material develop small needles of pale green amphibole, while some of the angular quartz fragments are penetrated by radiating needle-like clusters of a green mineral, too small for precise determination, but

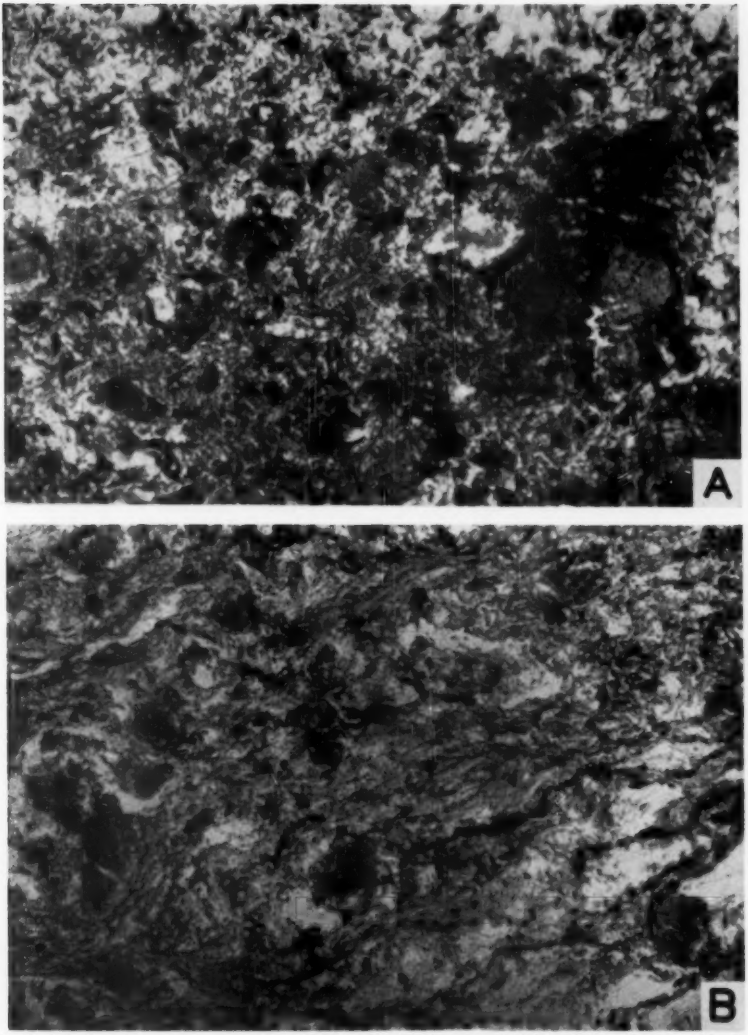
TABLE 1

	1	2	3	4	5	6
SiO_2	71.18	69.14	62.35	49.12	49.17	47.99
TiO_2	0.44	0.37	0.40	0.61	2.51	2.62
Al_2O_3	13.14	15.08	14.43	14.99	12.19	12.50
Fe_2O_3	1.05	1.51	1.25	0.97	1.10	3.40
FeO	2.37	0.84	3.76	9.21	10.88	11.12
MnO	0.06	0.03	0.04	0.07	0.29	0.31
MgO	2.25	1.08	3.40	6.58	6.83	6.11
CaO	1.48	2.79	5.97	10.65	10.76	12.16
Na_2O	4.31	5.31	2.78	2.58	2.47	2.98
K_2O	1.11	1.28	1.98	0.90	0.32	0.21
H_2O^+	1.91	1.96	3.65	4.11	3.58	0.44
H_2O^-	0.32	0.08	0.05	0.19	0.27	0.08
CO_2	0.15	0.00	0.07	0.00	0.02	nil
P_2O_5	0.08	0.12	0.00	0.12	0.00	0.00
	99.85	99.59	100.13	100.10	100.39	99.92
SG	2.70	2.91	3.02	3.15	3.20	3.47

1. Franciscan graywacke (302/74), Valley Ford (Bloxam, 1956, p. 493. *Anal.* E. H. Oslund)
2. Quartz-jadeite metagraywacke (302/138e), Valley Ford (Bloxam, 1956, p. 493. *Anal.* E. H. Oslund)
3. Quartz-glaucophane-lawsonite-schist (302/33), Valley Ford. *Anal.* T. W. Bloxam
4. Glaucophane-lawsonite-pumpellyite-schist (302/3), Valley Ford. *Anal.* T. W. Bloxam
5. Glaucophane-omphacite-garnet-schist (302/143b), Valley Ford. *Anal.* T. W. Bloxam
6. Eclogite (302/143c), Valley Ford. *Anal.* T. W. Bloxam

* Numbers refer to specimens in the Department of Geology, University of California, Berkeley.

PLATE 2

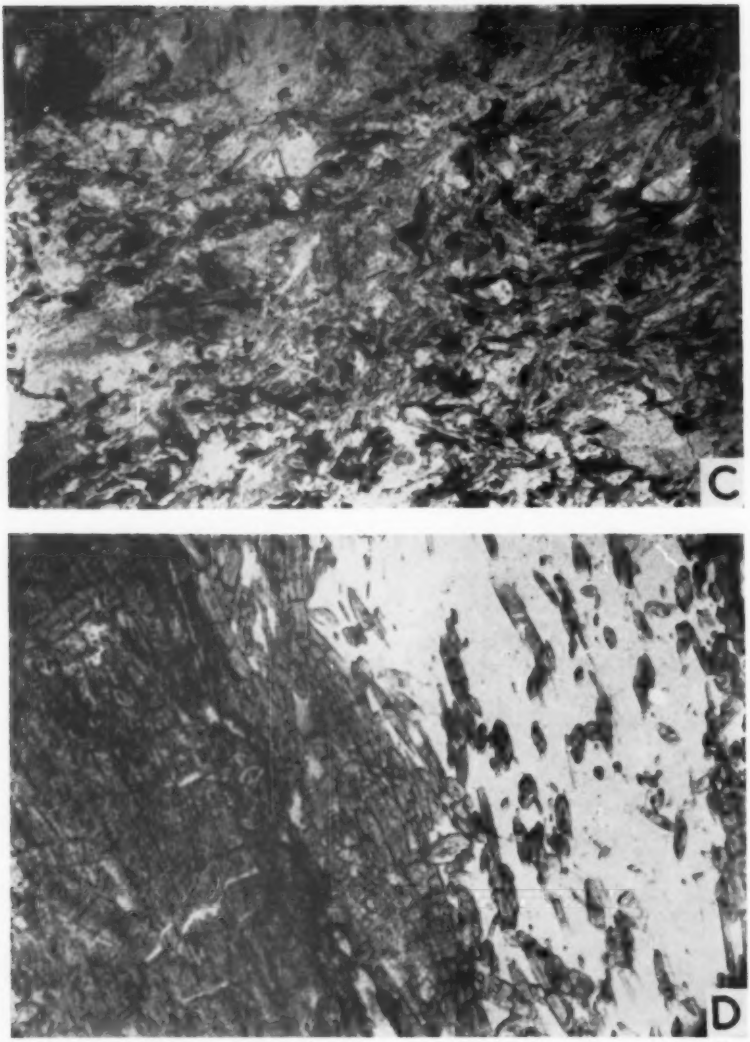


A. Jadeite-bearing metagraywacke (302/138), Valley Ford. The clastic texture of the graywacke is still preserved in this almost unfoliated specimen. Ordinary light ×8.

B. Quartz-(jadeite)-glaucophane-lawsonite-schist (302/33), Valley Ford. The specimen is a glaucophanized jadeite-bearing metagraywacke with turbid relicts of jadeite in a foliated web of glaucophane, lawsonite and muscovite. Ordinary light ×8.

whose habit resembles that of pumpellyite in the metacherts (p. 103). Patches of chlorite in the matrix occasionally develop an extremely narrow border of

PLATE 2

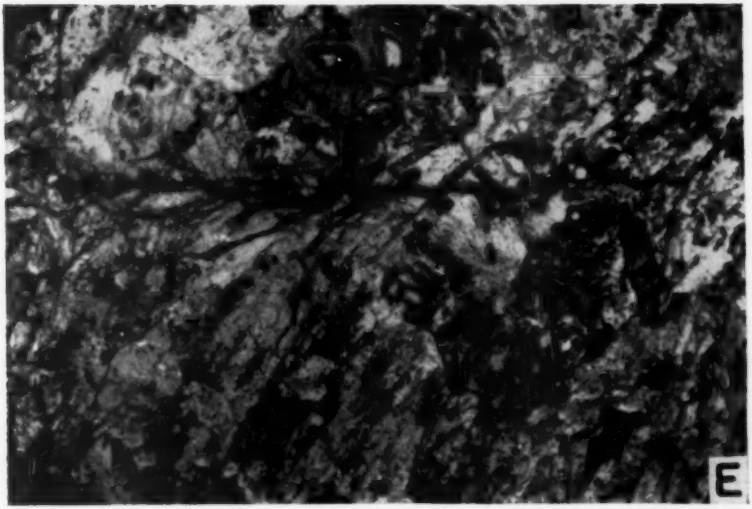


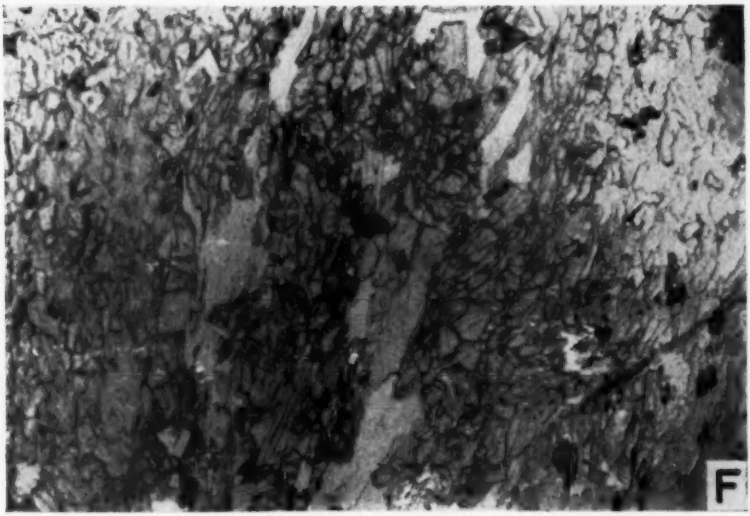
C. Glaucophane-lawsonite-pumpellyite-(muscovite, chlorite)-schist (302/3), Valley Ford. The rock is derived from a graywacke. Ordinary light ×8.

D. Metachert (302/57), Valley Ford. Contrasting bands of lawsonite (left) and quartz-glaucophane (right). Ordinary light ×8.

a faintly pleochroic fibrous blue mineral which may be glaucophane but, due to its small size, complete optical data could not be obtained.

PLATE 2





E. Altered eclogite (302/46), Valley Ford. Crystals of eclogite-garnet partially altered to lawsonite, pumpellyite and iron-ore. Stout crystals of glaucophane with crossitic margins are abundant, together with dark granular sphene. Ordinary light ×10.

F. Altered eclogite (302/24), Valley Ford. Glaucophane-omphacite-epidote-schist, the centre of the field being occupied by a thin schlieren of relict omphacite. Ordinary light ×3.

Metagraywackes and derived glaucophane-rocks.—The metagraywackes containing jadeite are confined to the area east of a fault which defines the line of a north-south trending creek. In a small waterfall near the head of the creek, and at various points downstream, extensive brecciation can be seen in the graywackes. The adjacent metagraywackes trend east-west with almost vertical foliation and enclose thin lenses of less altered graywacke, together with some altered and sheared igneous rocks (see below).

The jadeite-bearing metagraywackes from this, and other localities in California, have been described previously (Bloxam, 1956). They are the products of low-temperature metamorphism of graywackes and retain, to varying degrees, the clastic textures and minerals of the parent graywacke (table 1, 2 and plate 2A). In addition to jadeite (table 3, 1) the rocks contain varying amounts of glaucophane and lawsonite which generally define relict bedding and suggest derivation from more argillaceous bands. Larger blades of mica develop from an earlier sericitic mica (302/138).

A progressive increase in the amount of glaucophane and lawsonite can be observed in many of the basement metagraywackes (plate 2B). Glaucophane continues to crystallize from chloritic material in the groundmass and also at the expense of the jadeite which assumes a "dusty" appearance and exhibits marginal alteration to glaucophane (table 1, 3). In some cases the jadeite develops a pale green tint and increases in birefringence to 0.019, indicating a change in composition. The glaucophane is accompanied by idioblastic lawsonite and bands of mica. In some specimens (302/65) glaucophane is less prominent and jadeite crystals are imbedded in a groundmass of pale green, almost isotropic, chlorite.

Igneous rocks.—Thin lenses of variolitic basalt and tuff occur within the basement graywackes and metagraywackes. Those associated with the graywackes are composed of a brown glassy matrix charged with abundant iron-ore and riddled with microlitic needles of plagioclase (302/114). The plagioclase forms sub-radiate clusters and spherical aggregates; the rock strongly resembling the glassy variolites of Angel Island, San Francisco Bay (Ransome, 1894). Many of the variolites are brecciated and veined with quartz, albite and calcite; while others contain varying amounts of chert and argillaceous fragments, exhibiting transitions into tuffs.

Igneous rocks associated with the basement metagraywackes are of the same type but have been extensively sheared and metamorphosed. Abundant crystallization of pumpellyite is characteristic in these rocks. The pumpellyite occurs as a dense matt of small fibres which form a complex network of veins and infillings between the broken fragments of glassy variolite, but only rarely replace the fragments themselves (302/15). Varying amounts of glaucophane, lawsonite and late albite veins may also be present.

RESIDUAL MASSES

A characteristic feature of the area are masses, principally metamorphic rocks, scattered over the surface formed by the basement graywackes and metagraywackes. These masses vary in size from small boulders to crags 20 feet high (plate 1A). Their contacts with the underlying rocks are exposed in

some places and are perfectly sharp, suggesting that the masses cannot be precisely *in situ*.

The macroscopic fabric elements of the residual masses were plotted on a stereographic projection in order to evaluate their structure in relation to the basement rocks. However, no uniform structural pattern common to the various masses could be obtained, let alone any correlation with the structure of the basement rocks. Nevertheless, the distribution of the various blocks is not entirely random since the eclogites are confined to the south-east corner of the area, and the greatest concentration of chert and metachert occurs in the central region.

Considerable changes in both texture and mineralogy may occur within the compass of a single crag, making it possible to follow in some detail the relationships between the various metamorphic types.

The residual masses can be divided into the following groups: (i) chert, (ii) metachert, (iii) altered igneous rocks, (iv) eclogites, (v) glaucophane-schists, and (vi) epidote-amphibolites.

Chert.—The cherts range in color from red to dark green, the former being dominant, and are generally well bedded but shatter easily due to the development of numerous fractures normal to the bedding. The bedding is defined by thin leaves composed of an admixure of chert and shaley material.

In thin sections the cherts contain abundant radiolarian tests within a cryptocrystalline groundmass of quartz. Many are very impure (302/43), and contain finely dispersed chlorite stained with iron, together with much fine-grained unidentifiable subisotropic material. Recrystallization of quartz has generally occurred to varying degrees with resulting destruction of the radiolarian tests. In some cases the cherts have the appearance of micro-breccias; angular fragments of cryptocrystalline chert stained red with a fine limonitic dust being cemented by coarse vein-quartz (302/66).

Metacherts.—The earliest signs of alteration in the cherts (apart from recrystallization of the quartz) is the development of quartz-veins containing tiny stellate clusters of pumpellyite (302/67). At the same time the fine limonitic dust develops into larger aggregates with which are associated small sheaves of stilpnomelane. The crystallization of stilpnomelane is often accompanied by tiny curved fibres of riebeckite which riddle the quartz mosaic (302/50).

With the development of riebeckite the rock assumes a blue color which in some rocks alternates with flesh-pink bands rich in garnet (302/132). The garnet forms small idiomorphic crystals imbedded in the quartz mosaic, together with riebeckite and some crossite. An analysis of this garnet is given in table 2,1. from which it will be seen to approach spessartine in composition. Garnet-bearing metacherts of this type are common in the Franciscan and many similar rocks have been collected from Tiburon Peninsula (San Francisco Bay), and elsewhere.

Specimen series (302/57) collected from a single crag show transitions from brecciated chert with fractures filled with glaucophane and lawsonite, to banded quartz-glaucophane-lawsonite-schists. In the latter almost monomineralic bands of lawsonite alternate with bands of quartz-glaucophane (plate 2D).

In general the strongly banded and siliceous nature of the metacherts make them distinctive rocks in the field and, if one makes the not unreasonable assumption that all were derived from parent rocks of uniform highly siliceous composition, their origin is seldom in doubt.

Altered igneous rocks.—Blocks of altered variolitic basalts and tuffs are of frequent occurrence and are identical with the pumpellyitized variolites described from the basement metagraywackes (p. 102). Indeed, it is not always easy to determine whether some small blocks of altered variolite belong to the basement rocks or are residuals.

A few isolated boulders of carbonated serpentinite were observed among the residual masses but, in common with the latter, they cannot be regarded as precisely in situ. However, their possible significance will be considered subsequently (p. 108).

Eclogite.—Garnet-pyroxene rocks which are referred to eclogite are developed east of the fault where numerous large slabs rest upon the basement metagraywackes. Most of the eclogites exhibit varying degrees of deformation, but there are many exposures of fresh granular eclogite containing bright green pyroxene and evenly distributed red garnets (302/143). An analysis of eclogite from this locality is given in table 1, 6. In some exposures the relative proportions of pyroxene and garnet may vary within a distance of a few inches, so that pyroxene and garnet-rich phases develop. Muscovite and chlorite which are probably of secondary origin are always present, together with rutile and sphene. The chlorite has a rather distinctive radiate habit in thin sections and is a type characteristic of both the California eclogites and some glaucophane-

TABLE 2

	1	2		3		4	
SiO_2	37.39	38.01	Si	3.008		3.003	
TiO_2	0.42	0.62	Ti	0.048		0.023	
Al_2O_3	20.18	19.83	Al	1.915	1.992	1.844	2.019
Fe_2O_3	0.47	2.62	Fe‴	0.029		0.152	
FeO	5.47	23.68	Fe″	0.367		1.563	
MnO	25.96	0.81	Mn	1.770		0.052	
MgO	1.32	3.23	Mg	0.154	3.031	0.585	3.131
CaO	8.58	11.00	Ca	0.740		0.931	
	99.79	99.80					
						Almandine	55.93
						Pyrope	11.11
SG	4.14	4.04				Spessartine	1.81
n	1.798	1.801				Andradite	8.39
						Grossularite	22.75
							99.99

1. Spessartine garnet from metachert, Valley Ford. *Anal.* T. W. Bloxam
2. Garnet from eclogite, Valley Ford. *Anal.* T. W. Bloxam
3. Metal atoms in 1 on the basis of 6 oxygens
4. Metal atoms in 2 on the basis of 6 oxygens, and molecular composition

schists (Switzer, 1950). The pleochroism is from colorless to green, $N_\beta = 1.617$, $N_\gamma - N_\alpha = 0.006$, $2V_\gamma = 0°$ or very small.

The pyroxene is an acmitic diopside-jadeite or omphacite (table 3, 2) with variable optical properties, even in one thin section, and only average values can be quoted (table 4, 2). It is apparent that this green pyroxene is quite distinct chemically and optically from the colorless jadeite-rich pyroxene in the metagraywackes. No clear optical distinction can be made between pyroxenes referred to omphacite and many other aluminous pyroxenes (e.g. fassaite). Alderman (1936) gives the following optical data for the omphacites: $N_\beta = 1.67$-1.70, $Z:c = 36$-$42°$, $2V = 58$-$82°$ and SG = 3.30-3.36. However, the optics of many omphacites, including those from California, may depart from these values; particularly in respect to refractive index and extinction angle.

TABLE 3

	1	2	3		4			5		
SiO_2	60.64	53.12	53.31	Si	1.929			1.917		
TiO_2	1.29	0.31	0.26	Al	0.440	0.071	2.000	0.444	0.083	2.000
Al_2O_3	18.17	10.31	10.52			0.369			0.361	
Fe_2O_3	3.68	4.55	4.11	Ti	0.006			0.006		
FeO	0.53	3.00	2.84	Fe‴	0.078			0.112		
MnO	0.07	0.08	0.05	Fe″	0.089		0.991	0.086		1.016
MgO	1.09	7.54	8.92	Mn	0.005			0.002		
CaO	2.13	15.03	14.00	Mg	0.444			0.449		
Na_2O	11.36	5.81	5.90	Ca	0.561			0.552		
K_2O	0.25	0.03	0.05	Na	0.427		0.988	0.410		1.000
H_2O^+	0.78	0.13	0.16	K	—			0.038		
H_2O^-	0.11	0.06	—							
	100.10	99.97	100.12							
SG	3.27	3.30	3.34							

	6	7	8
Jadeite	77.68	26.26	28.38
Diopside	7.46	50.97	55.10
Acmite	12.39	12.92	11.95

1. Jadeite from metagraywacke, Valley Ford (Bloxam, 1956, p. 494. *Anal.* E. H. Oslund. Mineral contains 12.55% quartz inclusions)
2. Acmitic diopside-jadeite from eclogite, Valley Ford. *Anal.* T. W. Bloxam
3. Acmitic diopside-jadeite from eclogite, Healdsburg (Birch, 1943, p. 272. *Anal.* F. Gonyer)
4. Metal atoms in 2. on the basis of 6 oxygens
5. Metal atoms in 3. on the basis of 6 oxygens
6. Molecular percent end-members in 1.
7. Molecular percent end-members in 2.
8. Molecular percent end-members in 3.

Tilley (1938) has distinguished the omphacites from other aluminous pyroxenes of omphacite-type on the basis of SiO_2 (52-55.5%), and the coordination of the aluminum which, in the omphacites, largely replaces the Y group. From the analysis and structural formula for the Valley Ford eclogite pyroxene (table 3, 4), both SiO_2 and the co-ordination of aluminum are in agreement with omphacite.

An analysis of the eclogite garnet is given in table 2. 2 and compares with analyses of garnet from other California eclogites (Birch, 1943; Borg, 1956) although it differs in having slightly more Ca and lower Mg. The composition of this and other garnets from California eclogites and glaucophane-schists lies within, or very close to, the field of "eclogite garnet" (Eskola, 1921; Pabst, 1931, 1955; Borg, 1956).

Within a single exposure of eclogite the effects produced by deformation can be examined in some detail. The initial stages are largely mechanical and produce a crude flaser-structure defined by streaks of broken pyroxene and garnet. The development of glaucophane may be apparent macroscopically at this stage, but as the deformation increase blue streaks of the mineral become common. Within a few inches breakdown of the granular eclogite yields a crudely foliated glaucophane-rock with variously preserved remnants of eclogite garnet and pyroxene which form discontinuous lenses or schlieren. Some of the eclogites contain conspicuous radiating clusters of fibrous glaucophane, many over 2" long (302/143).

TABLE 4

	1	2	3
α	1.659 colorless	1.673 pale green	1.688 pale green
β	1.663 colorless	1.688 pale green	1.692 pale green
γ	1.670 colorless	1.696 green	1.710 pale yellow
$\gamma-\alpha$	0.011	0.023	0.022
$2V_\gamma$	86°	72°	70-82°
γ:c	53° (Y=b)	58° (Y=b)	44-50°
Disp.	Very strong	Moderate	Moderate (r>v)

1. Jadeite from metagraywacke, Valley Ford (Bloxam, 1956, p. 490)
2. Acmitic diopside-jadeite, Valley Ford
3. Acmitic diopside-jadeite, Healdsburg (Optical data from Borg, 1956, p. 1571)

Glaucophane-schists derived from eclogites.—As described above, glaucophane develops in association with the deformation of the eclogites. Further mineralogical changes can be observed in thin sections where the green eclogite pyroxene exhibits marginal replacement and pseudomorphism by glaucophane ($2V_a = 47°$, $Z:c = 4°$) in the form of moderately stout prisms which are often zoned, passing into deeper blue crossite ($2V_a = 0\text{-}30°$). The crossite may in turn have a marginal rim of blue-green hornblende (302/46). In some rocks, aggregates of small granular epidote crystals occur.

The eclogite garnet generally persists after the disappearance of the

pyroxene and, even in glaucophane-rocks where a strong superimposed schistosity and new minerals have developed, the distribution of the garnets, or their alteration products, preserve a vague remnant of eclogite texture. Mineralogical changes in the garnet are numerous and the mineral has been replaced, and even pseudomorphed by any, or more than one, of the following minerals: chlorite, lawsonite, pumpellyite, and occasionally glaucophane (plate 2E).

Rutile from the eclogite often develops rims of sphene which may be mantled by calcite. The subsequent history of the eclogite chlorite is not clear, but some at least has been observed altered to pumpellyite. Mica appears to be the only constituent of the eclogite that remains unaffected, although some recrystallises forming smaller crystals parallel to the schistosity. The precise composition of the mica requires further chemical investigation since the K_2O content of the eclogite and derived glaucophane-schist (table 1, 5 and 6) is lower than one would expect from the amount of mica present, assuming the latter to be muscovite.

The final mineral assemblage is typically represented by glaucophane/crossite-lawsonite-pumpellyite-mica-schists. Rocks of comparable mineralogy may also be derived from graywackes (below) so that in the absence of definite relict textures and minerals the origin of many glaucophane-schists at Valley Ford cannot readily be determined (plate 2F).

Glaucophane-schists derived from graywackes.—The development of glaucophane-schist from jadeite-bearing metagraywackes has been described from the basement rocks (p. 102). Similar rock-types occur as residual masses and one large crag may display various stages in the development of glaucophane and associated minerals (plate 2C).

Eventually the composition and mineralogy becomes similar to that of the altered eclogites, however the texture tends to remain finer in grain-size (table 1, 4). Glaucophane forms rather sinuous bundles of small fibres which are bent around idioblastic crystals of pumpellyite and lawsonite. Garnet has not been observed in these rocks (cf. Borg, 1956, p. 1572). There are some chemical differences between these rocks and those of definite eclogitic origin; in particular the minor oxides TiO_2 and MnO are consistantly lower (p. 111).

Epidote-amphibolites.—Rocks containing an amphibole other than glaucophane are common. Their mineralogy is relatively simple consisting essentially of epidote, hornblende and albite (302/23). Pumpellyite may occasionally be present in small amount, but lawsonite never occurs. Two types of amphibole have been observed; one strongly colored in thin section, and the other an almost colorless variety. The former is the same amphibole that occurs as rims or patches in association with glaucophane and crossite in the glaucophane-schists (p. 106). This mineral appears to be the same as "carinthine" (hornblende) described by Murgoci (1904) from the California glaucophane-rocks and has the following optical properties: X = colorless, Y = green, Z = deep blue-green, $N_\beta = 1.660$, $N_\gamma - N_\alpha = 0.024$, $2V_\alpha = 73\text{-}75°$ and Z:c = 17-20°. Hutton (1940) has described an actinolite ($Al_2O_3 = 3.88\%$) from albite-stilpnomelane-actinolite-schist in New Zealand which has optical properties very similar to the Valley Ford amphibole. The colorless amphibole has not been observed in association with glaucophane and has very weak pleochroism in

pale yellows and a $2V_a = 80\text{-}84°$, suggesting affinities with soda-tremolite or edenite.

Albite and epidote are always present, the former often in two generations, the first defining the foliation while the second forms later veins and irregular aggregates.

A few epidote-amphibolites contain relict glaucophane mantled by the blue-green hornblende. These rocks usually exhibit evidence of later brecciation and may represent retrograde glaucophane-rocks. However, there is no evidence indicating that all the epidote-amphibolites were so derived since many occur as lenses and schlieren within masses of glaucophane-schist against which they exhibit sharp boundaries.

MERCED FORMATION (PLIOCENE)

Horizontal or very gently dipping soft marine sands and clays of the Merced Formation form the higher ground around Valley Ford. The base of the Merced is exposed at one locality (see below) while elsewhere its base can be mapped with reasonable accuracy using topographic features. It is probable, however, that the present surface of Franciscan rocks corresponds very closely to the pre-Merced erosion surface.

Contact between the Merced and underlying rocks is exposed in the creek immediately below the bridge at the extreme southeast corner of the map. A vertical section of about 25 feet of Merced is exposed which is fine-grained throughout. A few blocks of glaucophane-schist not more than 1 to 2 feet long are imbedded in the Merced at its base but are quite absent at higher levels. Traverses over adjacent outcrops of Merced failed to reveal any incorporated blocks of glaucophane-schist.

ORIGIN OF THE RESIDUAL MASSES

The presence of numerous crags and boulders of Franciscan metamorphic rocks resting upon unaltered Franciscan graywackes indicates that the former cannot be precisely *in situ*. However, since the general metamorphic character of some of the residual masses conforms to that of the basement metagraywackes, and the distribution of certain rock-types is not entirely random (p. 103), the blocks do not appear to have undergone extensive transport. The enormous size of many of the blocks also tends to preclude large-scale transport.

It is probable that the blocks occupied their present positions before Merced times since many of the glaucophane-schist crags exhibit numerous *Pholas* borings (plate 1B) and horizontal wave-cut ledges produced by marine conditions; presumably the Merced sea. While several explanations are possible the writer is of the opinion that the residual masses formed an original cover of extremely hard and resistant metamorphic rocks, remnants of which remained as "collapsed" relicts upon a pre-Merced erosion surface which cut progressively lower into much softer graywackes and metagraywackes beneath.

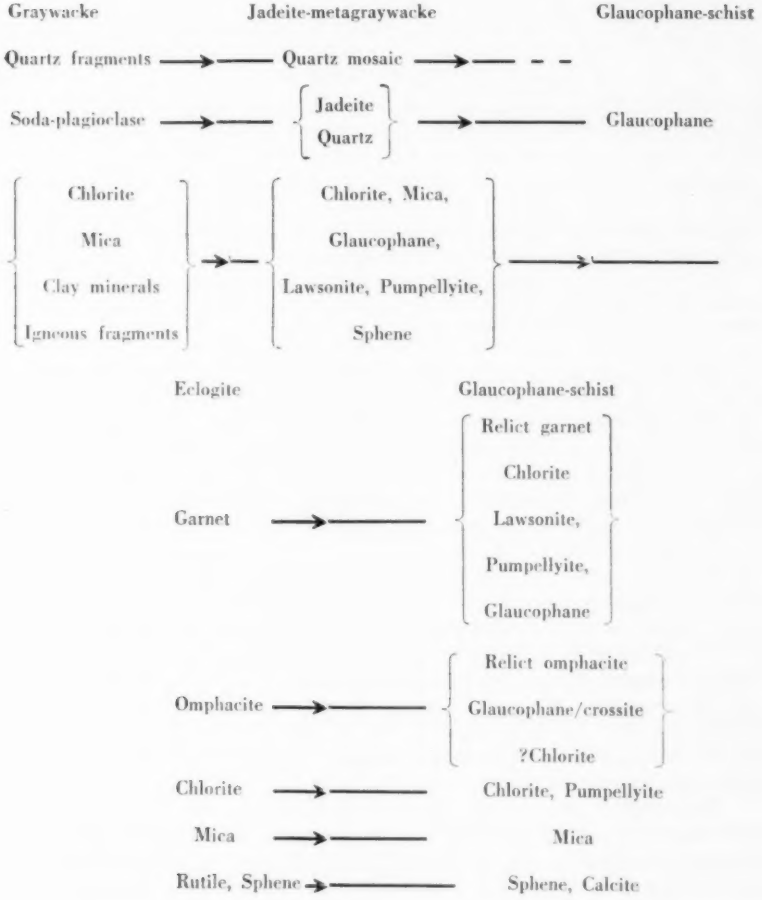
In California the glaucophane-schist metamorphism may occur at various levels in the thick geosynclinal belt of Franciscan rocks. The metamorphism is frequently, but not exclusively, associated with subjacent bodies of serpentinite (Ransome, 1894; Taliaferro, 1943). While no serpentinite occurs *in situ* at

Valley Ford, a few blocks of carbonated serpentinite (p. 104) may represent remnants of former ultrabasic intrusions.

CHEMICAL AND MINERALOGICAL RELATIONSHIPS

From an examination of the textural and mineralogical features of the metamorphic rocks two modes of origin are indicated; the first from graywackes, cherts and igneous rocks, and the second from eclogites. The abundance of the latter type indicates that eclogite must have been originally present in considerable amounts (Borg, 1956). The observed mineralogical relationships and transformations are summarized in table 5.

TABLE 5



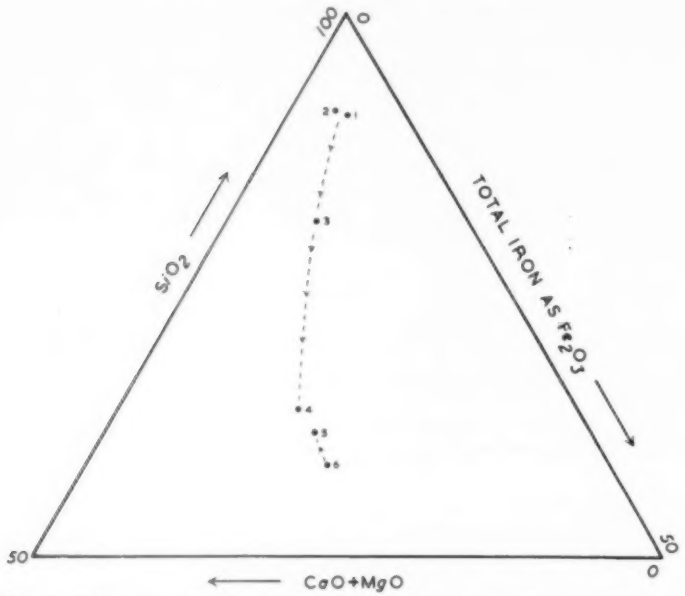


Fig. 2. Triangular diagram illustrating the major changes in composition during glaucophanization of the graywackes and eclogite at Valley Ford. The analyses plotted correspond to the numbers in Table 1.

All the rock analyses have been plotted on a ternary diagram (fig. 2) in terms of SiO_2, total iron as Fe_2O_3 and $MgO + CaO$; these showing the greatest variation among the major oxides. There is virtually no change in bulk composition during the development of quartz-jadeite metagraywacke from graywacke. Hence jadeite forms under conditions appropriate to the glaucophane-schist facies in rocks of rather restricted chemical composition (low Fe, Mg and Ca). The succeeding phases in which more glaucophane develops, some at the expense of the jadeite, are accompanied by an increase in Fe, Mg, Ca and a decrease in Si and Na. This trend continues towards a final assemblage which is commonly glaucophane-lawsonite-pumpellyite-schist. These changes in bulk composition could be due either to differences in the original composition of the graywackes, or to metasomatism. While the former possibility cannot be entirely excluded, the extremely constant chemical composition of the Franciscan graywackes (p. 98) appears to be reflected through analyses 1-4, table 1; particularly TiO_2, Al_2O_3 and MnO. This, in conjunction with evidence from relict textures and transitional phases suggests metasomatism rather than large differences in original composition.

The initial chemical composition of the eclogites must be one which is more appropriate to the development of glaucophane-schist since only small changes in composition are involved in their transformation; the final product approaching the composition and mineralogy of the glaucophanized gray-

wackes. Of the minor oxides, both TiO_2 and MnO are consistently higher than in the metagraywackes.

The absence of albite in the greater majority of glaucophane-rocks at Valley Ford, and the progressive elimination of omphacite and eclogite garnet, indicates that these minerals are unstable in the glaucophane-schist facies. However, manganese garnets are apparently stable in association with glaucophane (p. 103; Miyashiro and Banno, 1958). The eclogites must therefore have formed prior to, and under different conditions, from those of the glaucophane-schist facies (Turner, 1951). In general the mineralogical character of the glaucophane-schists has affinities with the greenschist facies (Turner, 1951; Miyashiro and Banno, 1958), but the instability of albite and the presence of hydrous minerals with high density, notably glaucophane, lawsonite and pumpellyite, suggests certain conditions perculiar to the glaucophane-schist facies, possibly high water-vapor pressure (Yoder, 1950).

ACKNOWLEDGMENTS

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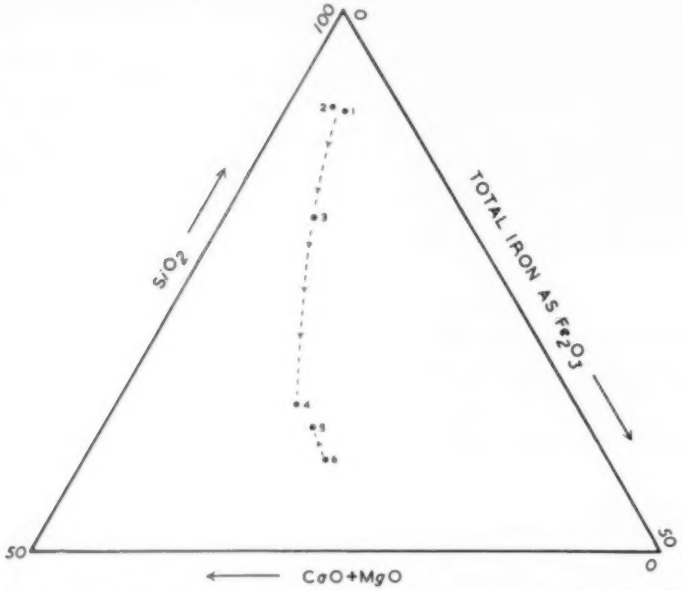

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TUFFISITES AND MAGNETITE TUFFISITES FROM TORY ISLAND, IRELAND, AND RELATED PRODUCTS OF GAS ACTION

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ABSTRACT. On Tory Island an unrecorded small swarm of saccharoidal leucocratic aplitic veins invades the 'Older Granite' of Donegal. Microscopically the veins have granophyric margins; intimate shattering and penetration of the granitic country rock is accompanied by incorporation of xenoclasts of granite and neighbouring meta-sediment. One vein is replete with magnetite grains (making it completely black in hand-specimen), which crystallized between the two phases of intrusion of this composite aplite. Detailed structural and textural features suggest these veins are tuffisites resulting from gaseous fluidization (gas streaming). The magnetite appears to derive from decayed granitic mafics. The magnetite has cell dimension of pure Fe_3O_4 (Curie temperature 585 ± 5°C) and a very low Ti-content; this implies low temperature crystallization according to Buddington, Fahey, and Vlisidis (1955), but the universal validity of their hypothesis is questioned. Evidence for very high temperature crystallization of these tuffisites is advanced (c.f. Slieve Gullion tuffisites described by D. L. Reynolds). Xenolithic and xenocrystic rocks in numerous environments, and sedimentary dykes associated with igneous dykes, are discussed in relation to positive evidence of violent gas-phase activity during intrusion processes.

INTRODUCTION

Tory Island lies in the Atlantic Ocean six miles north of the most northwestern tip of Ireland. Its eastern extremity comprises massive Dalradian quartzite, whilst the major portion of the island is younger Caledonian granite. A well-developed peneplain bevels the island and now dips to the south-southwest; this results in cliffs over 200 feet high in the northeast, and broad tidal reefs in the southwest.

The granite of Tory Island is part of the 'Older Granite', the oldest member of the complex composite Donegal Granite. The 'Older Granite' extends some 23 miles south of Tory, although it is invaded by several younger granites (see fig. 1). As a result of the discussions of pioneers like Grenville Cole, much of the Donegal Granite is classic ground. Recently all the 'Older Granite' has been re-examined and partially described by Gindy (1953), Pitcher (1953a, 1953b), Whitten (1957a, 1957b), and Read (1958).

The literature on Tory Island is very meager. Giesecke (1826) recorded a fleeting unpleasant visit during which he collected a few granite specimens. Haughton (1862, 1871) and Scott, Griffiths, and Haughton (1863) merely gave a single analysis of Tory granite. The island is included on the one-inch Sheet 3 of the Geological Survey of Ireland; the chapter on the island in the accompanying memoir (Hull, et al., 1891) deals almost exclusively with archaeology and legendry history. Whitten (1957a, 1957b) recently described the broad geological structure of the island and the granite-quartzite contacts.

The Tory Island aplitic veins, which will be shown to be tuffisites, have not been mentioned previously. They traverse the coarse-grained granite of

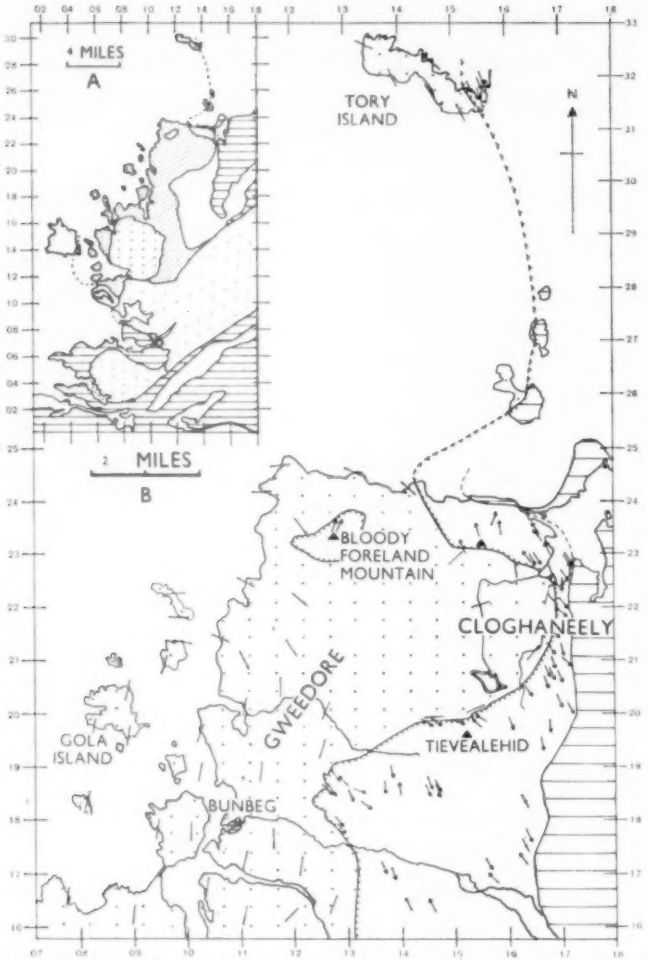


Fig. 1. Geology of northwest Donegal, Ireland.

A. 'Older granite' (close stippled), and Ardara (080045), Rosses (145090), and Main (150100) 'Younger granites'. Dalradian quartzite (unornamented), and pelitic and calc-silicate rocks (ruled).

B. Northern part of the 'Older granite' showing trend of foliation in granite, and plunge of quartzite lineations.

Morard; this hill appears to be about three-fourths of a mile west of the granite-quartzite contact (fig. 2).

THE GRANITE

The granite is coarse- to medium-grained biotite hornblende granodioritic

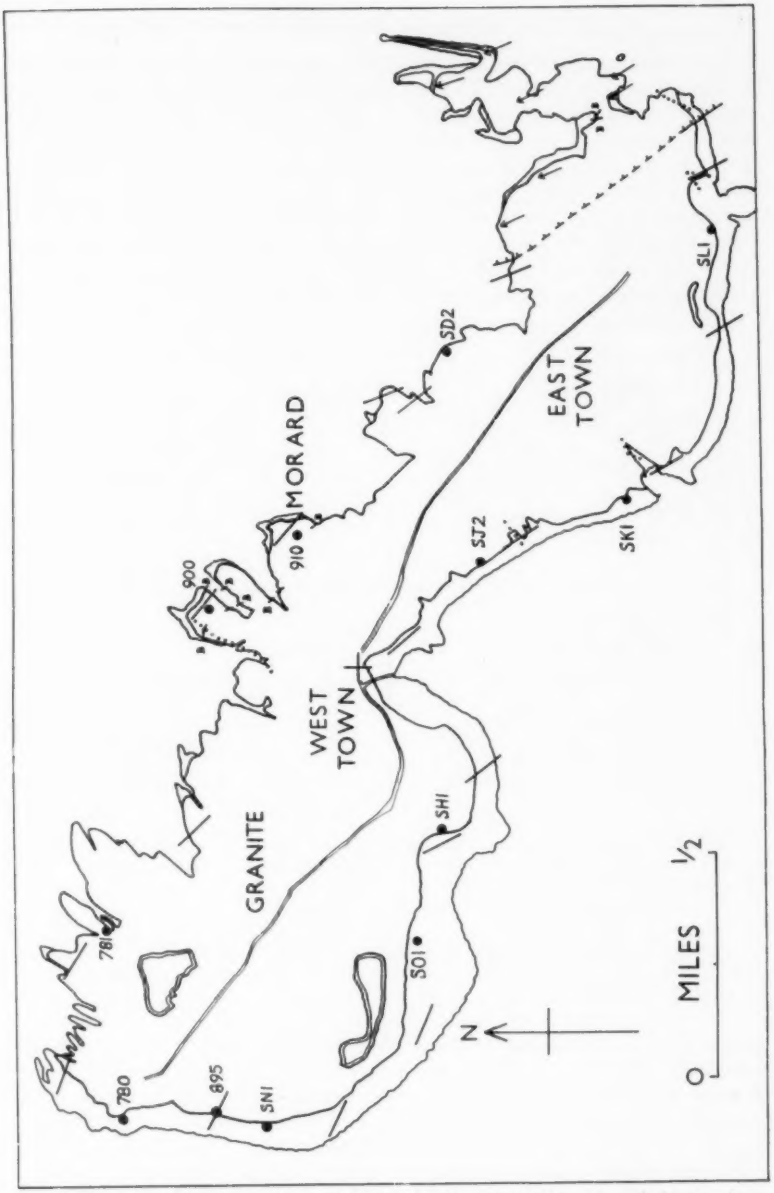


Fig. 2. Tory Island with low and high tide shorelines indicated, and showing locations of specimens quoted in table 1.

Lines = trend of foliation in granite; Arrows = plunge of lineation in quartzite; TTTT = granite-quartzite contact; .... = microgranite dykes; B = Tertiary basalt/dolerite dykes.

or adamellitic rock. The modal composition and texture are more variable than on the adjacent mainland (Whitten, 1957b). For example, microcline microperthite porphyroblasts are variable in size and abundance; towards the west and southwest they are more euhedral and larger, often reaching 1½ inches in length. Also, compared with the mainland, modal variation is rather large (see table 1 and fig. 3), although probably coarseness of grain-size accentuates the apparent variability of the quartz and feldspars. The granite in Tory is relatively close to the contact, and, particularly at the eastern end of the island, it contains numerous quartzitic xenoliths ranging from considerable rafts (Whitten, 1957a) to micro-xenoliths. The latter are easily recognized in thin-

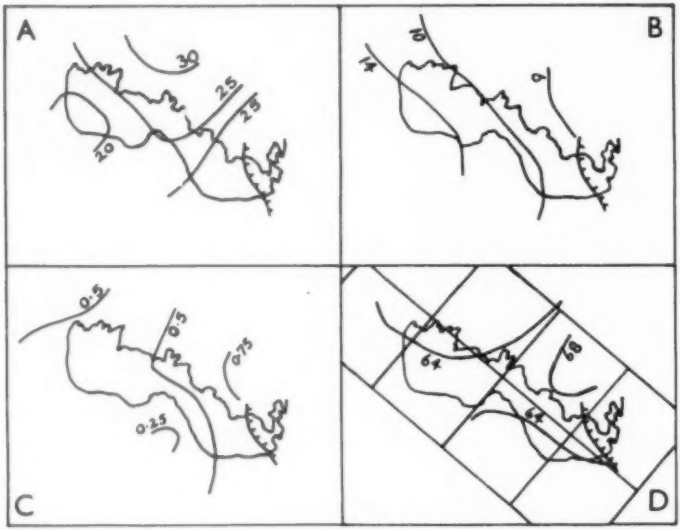


Fig. 3. Modal variation of granite based upon 12 values in table 1. Isopleths drawn for average values in each 1 square mile area.

A. Total modal quartz percentage.

B. Colour index.

C. Ratio of microcline microperthite to plagioclase.

D. Total modal feldspar percentage. The 1 square mile grid squares are indicated.

section by their distinctive texture. Other enclosures within the granite are restricted to the ubiquitous autoliths, except for a large epidioritic mass (some 60 yards across) on the coast southwest of West Town. Autoliths occur sparingly throughout the granite, and are usually 6-12 inches across. They are angular, irregular, rounded, or discoidal and thus less consistantly discoidal than in most of the 'Older Granite'. The largest dimension tends to lie within the subvertical foliation of the granite. The autoliths occasionally contain feldspar porphyroblasts.

The granite is cut by a number of dissimilar microgranite dykes, which

TABLE 1
Modes of Granitic Rocks of Tory Island, Co. Donegal

Specimen No.	781	900	910	SD2	780	895	SH1	SJ2	SK1	SL1	SN1	SO1
Quartz	25.1	41.0	24.1	25.2	20.8	17.3	20.8	23.3	25.4	29.9	22.6	15.6
Plagioclase	52.3	31.8	36.1	42.2	42.3	47.1	50.6	53.2	47.8	37.7	48.3	48.7
Microcline	14.2	21.7	31.6	29.1	22.0	21.0	15.4	12.2	11.1	25.0	17.1	17.1
Biotite	6.6	4.8	6.6	2.8	9.0	7.1	8.7	8.0	10.6	5.0	8.3	9.6
Hornblende	1.7	0.4	1.2	0.4	5.2	7.1	3.9	2.8	4.2	1.7	3.4	8.5
Sphene	0.06	nd	nd	—	nd	nd	0.1	0.1	0.2	0.2	—	0.3
Magnetite	0.06	nd	nd	0.1	nd	nd	0.2	0.3	0.2	0.4	0.05	0.2
Apatite	—	nd	nd	0.08	nd	nd	—	0.05	0.09	0.04	0.2	0.05
Others	—	0.2	0.4	0.08	0.6	0.4	0.2	0.1	0.3	0.09	0.1	—
Chayes' Classification	P_{13}	—	$(321)_8$	$(321)_2$	$(321)_{15}$	—	P_{13}	P_{13}	P_{13}	$(312)_7$	P_{13}	—

nd = not differentiated in analysis, but included in 'Others'.

Chayes' Classification according to Chayes (1957), and each value should be preceded by $\frac{h}{h}$ II

Specimens with numerical identification quoted in part in Whitten, 1957b (p. 255).

Specimens with identification numbers preceded by 'S', supplied by Mr. Dick Savage of Tory.

presumably belong to the same cycle of activity as the granite. Occasional Tertiary basaltic dykes are to be seen in the cliffs.

THE TUFFISITE APLITES

On Morard the coarse-grained granite (specimen 910, table 1) exhibits a weak foliation containing a lineation trending at 133 (i.e. 133° E. of N.). The dominant joints trend at 015, 139, and 262. The granite of Morard is traversed by a swarm of aplitic dykes which are slightly sinuous and sometimes cross one another. Their general trend is 028. Although aplitic in hand-specimen, these rocks are seen to be typical tuffisites (Cloos, 1941) in thin-section.

These pink dykes show a saccharoidal texture and are usually two or three inches wide. Seen on a cut surface the texture is slightly banded (reminiscent of the aplitic microgranite of Slieve Gullion figured by Reynolds, 1951, Plate I, fig. 4) and less homogeneous. Apart from very occasional small clusters of biotite, the rocks are wholly leucocratic. The average grain-size is 0.02 inches, but some patches are uniformly finer (0.004 inches or even less). Microcline microperthite slightly exceeds oligoclase, whereas quartz occupies some 15-20 percent. However, it is impossible to effect a point counter analysis of this material because it is crowded with both xenolithic and xenocrystic enclosures.

The xenolithic material comprises both quartzitic and granitic fragments. The former are shapeless and up to 0.25 inches long. Their texture is typical of the regional Dalradian quartzites, but it has sometimes been slightly 'simplified', like the quartzite micro-xenoliths within the 'Older Granite' near Bloody Foreland (see Whitten, 1957a).

The granitic fragments are dominantly aggregates of plagioclase with a little interstitial quartz; these are not synneusis clots because the adjacent crystals interlock. Not more than three or four plagioclase crystals (0.05 – 0.15 inches long) occur in these xenoliths. Identical individual euhedral and subhedral plagioclase crystals occur as xenocrysts. Apart from their size, the advanced state of decomposition (which emphasizes the zoning) immediately differentiates these xenoliths from the plagioclase of the aplitic groundmass; the latter is quite fresh, or lightly dusted with 'white mica'. As a result of corrosion and replacement, the peripheral zone of the xenocrysts is sieved by rounded or bulbous quartz grains (see fig. 4a). The xenocrysts are occasionally traversed by dendritic areas of recrystallized plagioclase which have been cleared of decomposition products. These areas are apparently in optical continuity with the host grain, although possessing a different extinction angle (fig. 4b).

Microcline xenocrysts are less obvious, but the aplitic groundmass encloses numerous largish angular fragments of microperthite, apparently derived from the granite. These crystals are considerably kaolinized, and carry a little secondary 'white mica'.

THE MAGNETITE TUFFISITE APLITE

Associated with these 'aplites' is one rich in magnetite; it is one inch wide, can be traced 2 yards, and dips at some 25° west. In hand-specimen this unusual vein appears to be composed of a dense homogeneous schorl-like sub-

stance, studded along its center with quarter-inch angular pink feldspars. The vein is penetrated by thin very fine-grained aplitic stringers which are sub-parallel to its walls. The precise relationships vary rapidly along the length of the vein.

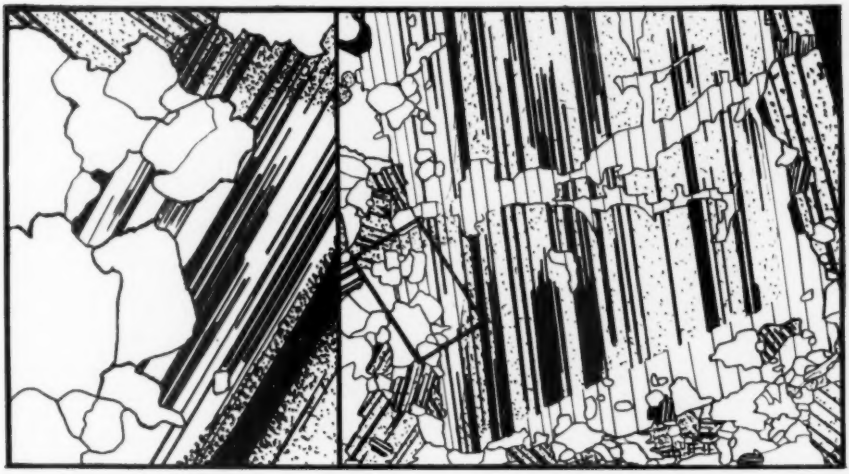


Fig. 4. Detail of plagioclase xenocryst, aplitic tuffisite (spec. 911), Morard, Tory Island. Plagioclase: symbolic albite twinning; Quartz: unornamented; Microcline: cross-hatched; Magnetite: black. Camera lucida drawings; scales = 0.01 inch.

A. (left) Marginal corrosion of xenocryst by quartz.

B. (right) Xenocryst traversed by dendritic veins of recrystallized plagioclase; these veins are in optical continuity with the host, and are cleared of decomposition products, and, like the marginal zone penetrated by quartz, show different extinction angles from the host. Inset shows location of fig. 4A.

Thin-sections of different specimens of this vein are of considerable petrogenetic interest. The broad features are similar to those of the leucocratic aplites, except that this vein is replete with tiny granules (often euhedral octahedra) of magnetite. The magnetite always occurs between the crystals of the quartzo-feldspathic main aplite; though obviously developing at the expense of the latter, magnetite is never enclosed within the leucocrates. The numerous xenolithic and xenocrystic enclosures are free from magnetite (save sometimes the extreme margins), and there is no special accentuation of ore at their periphery. Figure 5 shows the main aplitic vein cut by later magnetite-free veinlets margined by especially magnetite-rich zones.

In hand-specimen this vein is very clearly defined, but the aplitic material is difficult to delimit in thin-section. The granitic country rock flanking the vein is highly shattered and penetrated by innumerable aplitic veinlets. Such veinlets are sinuous and irregular, or straight and parallel-sided; they sometimes follow crystal boundaries, but more frequently pass randomly outwards

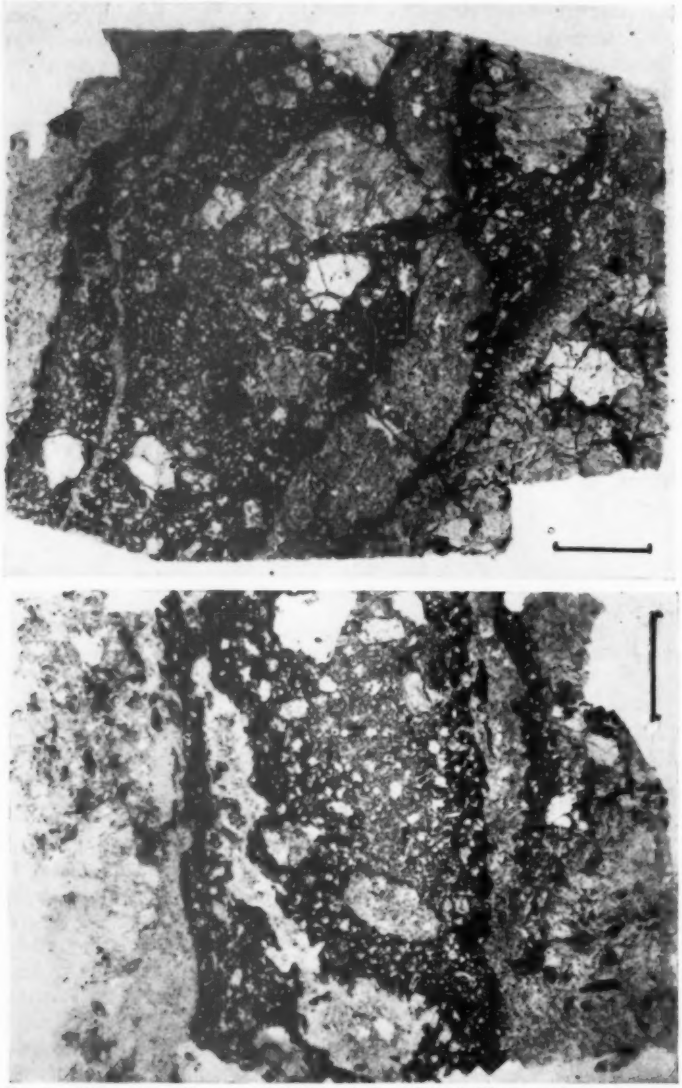


Fig. 5. Magnetite tuffisite vein of Morard, Tory Island. Photographs (ordinary light) of thin sections; scales = 0.2 inch.

A. (above) specimen 907. B. (below) specimen 908.

through the crystals of the granite. Sometimes a crude zonation flanking the main vein can be detected which reflects the degree of permeation of the

granite by veinlets. Marginal to the main vein, and, on a smaller scale, marginal to the veinlets, plagioclase recrystallizes and is cleared of the kaolinite and 'white mica'. The tiny veinlets sometimes swell out irregularly to give aplitic patches within the granitic wall rock. Some of the large veinlets (0.02 inches wide) are flanked by granophyric selvedges up to 0.005 inches broad.

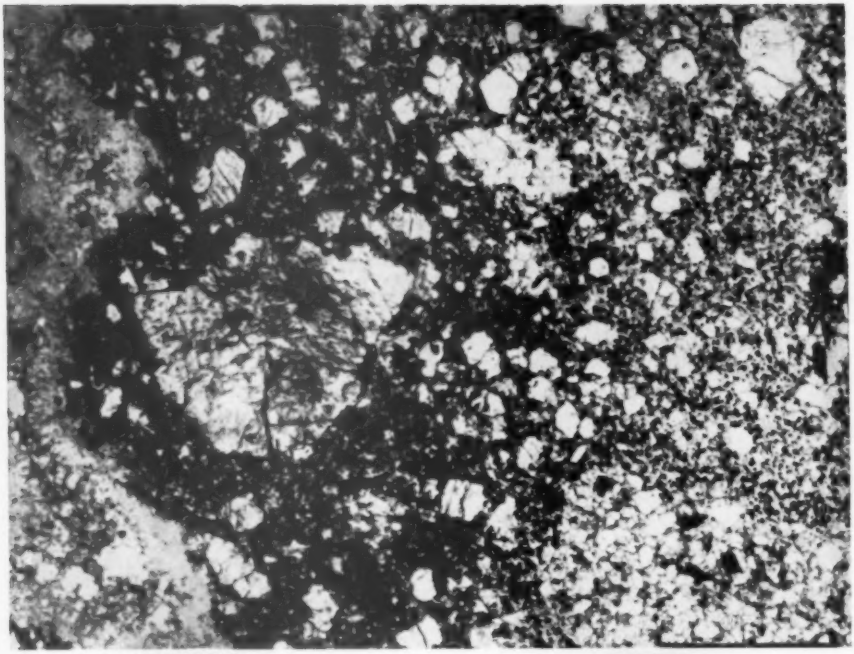
Hornblende and biotite in the granite immediately adjacent to the aplites are considerably altered. Brown biotite disintegrates into shredded red-brown whisps, and when actually caught in the magnetite aplite, these shreds tend to be dusted with minute magnetites. In some cases relics of biotite are enclosed within grains or aggregates of magnetite. Cores of subhedral hornblende crystals are replaced by limonite, magnetite, or a mixture of the two. The peripheral amphibole recrystallizes as a large number of disorientated crystals, whose grain-size approximates that of the enclosing aplite. When incorporated in a major aplitic vein the amphibole becomes extremely anaemic in color, almost non-pleochroic, and each crystal unit develops a diablastic texture (the enclosures being of quartz).

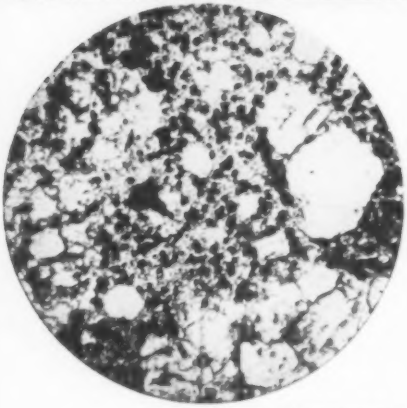
Each aplitic veinlet within the magnetite aplite has a well developed granophyric margin. The structure inside is more equi-granular, with a grain-size of 0.0025 – 0.005 inches, and xenocrysts are uncommon. Through the constriction in the large veinlet of figure 5a, and beyond the tip of the magnetite zone of figure 7a, the aplite develops a 'fluxional' structure due to a habit orientation of the crystals, which are locally 0.00012 – 0.00004 inches in diameter. The overall texture of these veinlets very closely resembles that of the composite tuffisite (aplitic microgranite) vein in Adavoyle quarry, Foughill, figured by Reynolds (1951, Plate II, fig. 6). Reynolds claimed that the curvilinear forms shown in her illustration are reminiscent of glass shards. In the veinlets, curvilinear forms, composed of quartz crystals with unusual habit, are rather more obvious; marginal granophyre is also better developed than at Adavoyle.

THE DISTRIBUTION OF THE MAGNETITE

The magnetite is usually euhedral and always occurs between the aplitic quartzo-feldspathic crystals (fig. 6b). The straight edges of the magnetite have developed at the expense of pre-existing leucocratic minerals. The leucocratic xenoliths and xenocrysts never contain magnetite, although, as described above, magnetite is associated with the rare xenocrystic amphiboles and micas.

Occasional haematitic veins run right across a thin-section sub-parallel to the main vein walls, traversing both xenolithic and groundmass material. However, magnetite grains consistently cut the haematite veins. It follows that after the stabilization of the aplite and its incorporated xenoliths the magnetite grew, in part by replacement. This growth was accomplished while magnetite gained entry neither into the xenocrysts nor into the fine-grained aplitic xenoliths (figs. 5-7). An exception is the xenocrystic and wall-rock (granitic) plagioclase which was recrystallized and to a limited extent 'invaded' by iron ore. Here, a sharply defined zone of feldspar develops a more albitic composition, and is cleared of decomposition products. The cleared zones are often crescentic, being better developed where magnetite is more abundant within the





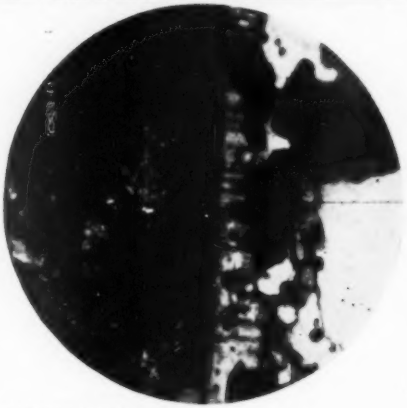


Fig. 6. Magnetite tuffisite specimen 908 (fig. 5B).

A (above). Photograph (ordinary light) of thin section showing concentration of magnetite marginal to the tuffisite veinlet. The veinlet has distinct fine grained granophyric margins. Xenocrysts and xenoliths (which are free from magnetite) abound in the magnetite tuffisite. Scale = 0.067 inches.

B (left). Photomicrograph (plane polarized light) of teh magnetite aplitic tuffisite. Diameter of field = 0.6 inches.

C (right). Photomicrograph (crossed polars) of the margin of the main magnetite tuffisite against the granitic country rock. The large plagioclase is part of the granite; it has recrystallized marginally and developed a diablastic texture due to invasion by magnetite and quartz. Diameter of field = 0.12 inches.

peripheral feldspar. Albite twinning is not always continuous across the junction of the border zone. Removal of the decomposition products from the marginal zones did not quite keep pace with the advancing marginal corrosion, because, under x400 magnification, trails of decomposition products are seen which do not stop abruptly at the junction. The outer margin of such feldspar is very ragged in contact with aplite; this is not an original feature, as evidenced by feldspars protruding from the granite and truncated by the veins (fig. 6c). The fringe of magnetite-enriched feldspar often diablastically encloses minute quartz blebs (fig. 6c).

The concentration of magnetite along the flanks of the aplitic veinlets (figs. 5 and 6a) presents a petrogenetic problem. If the magnetite were a late introduction it could be argued that the veinlet acted as a barrier against which the ore-producing media accumulated causing excess magnetite growth. Since the xenoliths have no similar ore mantle (fig. 6a), this is probably not the case. If the ore were not post-veinlet, the veinlets must have pierced magnetitic aplite, and their extraordinary shapes prove they are not dilational structures. Material in space now occupied by veinlets has been either removed or reconstituted—the iron certainly has been removed, and it has probably been driven into the adjacent host aplite. Xenoliths in the peripheral aplite still remained immune to invasion by magnetite, but the flanks (especially on the veinlet side) of these resistant enclosures have been more heavily enriched (figs. 5a and 6a). Margins of veinlets are usually knife-sharp, but sometimes a granule or two of magnetite protrudes about 0.007 inches into the marginal granophyre from the magnetite-enriched main aplite; clearly retreat of magnetite from developing veinlets has not kept pace with advance and scouring of the aplitic material.

These late aplitic veinlets are just as common adjacent to the ore-bearing vein, but they are difficult to detect except by microscopic examination under crossed polars.

THE COMPOSITION OF THE MAGNETITE AND ITS SIGNIFICANCE

Professor J. Verhoogan and Mr. L. Van de Griendt of the University of California kindly examined the Tory Island magnetite. They report that it contains about 5 percent haematite, which occurs as lamellae on (111) planes of the magnetite. The majority of these lamellae occur near, and extend to, the crystal boundaries, suggesting oxidation rather than exsolution. The cell dimension of the magnetite is 8.396—that, of synthetic pure Fe_3O_4. The Curie temperature is 585 $\pm$ 5°C, which suggests the TiO_2 content is probably less than 0.3 percent.

In an extensive review of natural magnetite, Buddington, Fahey, and Vlisidis (1955) claimed the TiO_2 content drops with decreasing temperature of formation. Their examples with less than 1 percent of TiO_2 are included in table 2. They (Table 10, p. 518) suggested that 1-2 percent TiO_2 may be equated with temperature limits of formation of 550-400°C (late stage granitic pegmatite and low temperature amphibolite facies associations). As the authors indicated, and Heier (1956) emphasized, numerous factors, such as pressure and variation in chemical composition of ore-forming solutions, militate against

TABLE 2

Magnetites with Low TiO_2 Content

	TiO_2%
Specimens from United States of America	
A. *Adirondacks* (Buddington, Fahey, and Vlisidis, 1955, p. 508)	
1. Muscovite microcline granite pegmatite. Spec. 210-B	0.51
2. Hornblende granite gneiss; low range of amphibolite facies; microcline and plagioclase as separate grains, sphene present	0.64
Specs. 12 and 188	0.70
B. *Alaska* (Zies, 1924, p. 176; Grantz, 1956)	
3. Fumarole 148, Valley of Ten Thousand Smokes, 1919	0.005
4. Pyrometasomatic replacement, Tuxedni Bay, Spec. C.	*ca.* 0.004
C. *Minnesota* (Gruner, 1934, p. 757)	
5. Cement of 'ore conglomerate', Mesabi Range.	< 0.01
D. *New Jersey* (Buddington, Fahey, and Vlisidis 1955, p. 509)	
6. Pyroxene syenite gneiss reconstituted in amphibolite facies. Spec. 147	0.98
7. Microcline (slightly perthitic) granite pegmatite. Spec. B-193	0.89
E. *Rhode Island* (Buddington, Fahey, and Vlisidis, 1955, p. 511)	
8. Westerly pink biotite microcline oligoclase granite. Spec. 221	0.89
F. *Utah* (Wells, 1938, p. 486)	
9. Iron ore veins cutting monzonite	see Table 3
Specimens from Canada	
G. *Nova Scotia* (Harrington, 1907)	
10. Veins in trap-rocks, Digby, Annapolis County	0.24
Specimens from Scotland	
H. *Isle of Skye* (Tilley, 1951, p. 662)	
11. Magnetite-chondrodite-skarn with diopside	0.18

the TiO_2 content of magnetites being used as a precise geological thermometer. However, Buddington (1956, p. 515) " . . . maintained that the temperature control is so much a greater factor where the iron-titanium oxides are essentially ilmenite and titaniferous magnetite that the composition of magnetite belonging essentially to the system Fe_3O_4—$FeO.TiO_2$ *does* have value as an approximate geologic thermometer if the usual discretion necessary in most petrological interpretations is exercised."

In view of this hypothesis the distribution of magnetite with low TiO_2 content would be of interest; unfortunately, few relevant analyses have been published, but attention may be drawn to the following examples.

Magnetite lining fissures vents of fumarole 148, Valley of Ten Thousand Smokes, Alaska contained 0.005 percent TiO_2 (Zies, 1924, p. 175-176). The magnetite was derived through vapor phase activity, and in 1919 steam was escaping at 239°C (Zies, 1924, p. 167; 1929).

0.44 percent TiO_2 in magnetite of Tuxedni Bay, Alaska (see table 2) is probably a slightly low value as the analyzed material was a grab sample of high-grade portions of a massive magnetite lens. However, the true value must be <0.5 percent. "The association of magnetite with andradite-rich garnet, epidote, diopside, chlorite, plagioclase, idocrase, and a trace of iron and copper sulfides . . . is characteristic of magnetite deposits of pyrometasomatic origin" (Grantz, 1956, p. 105).

Wells (1938) thought sudden release of pressure and liberation of $FeCl_3$ and H_2O gas from monzonite stocks still at a high temperature, resulted in some magnetite-haematite veins cutting monzonite in Utah. These gases found ready egress to the surface along tension fissures without appreciably heating the country rocks; the general abundance of volatile components is proved by the abundance of tuffs and ignimbrites. The TiO_2 content of the magnetite is very low (see table 3).

Tilley (1951) described the metasomatism at a granite-dolomite contact in Skye; in the second phase of skarn production, subsequent to the consolidation of the marginal granite, magnetite-chondrodite skarns with diopside developed. The magnetite contains 0.18 percent TiO_2.

Gruner (1934) described the magnetite cementing some ore conglomerates of the Mesabi Range as the purest ever analyzed. The analysis showed

FeO	30.94
Fe_2O_3	69.05
Mn	trace
	99.99

Gruner claimed this magnetite to be of hydrothermal origin.

In this section a brief review has been attempted of some TiO_2-poor magnetites. Definite evidence of temperature of each paragenesis is lacking, but the associations do not seem to imply the universally low crystallization temperatures claimed by Buddington, Fahey, and Vlisidis (1955). In fact, some of the associations could imply fairly high temperatures. Many of the parageneses are associated with late-stage igneous activity, and in some cases, including that described by Gruner (1934), it is likely that the composition of the magnetite was essentially controlled by the elements available (including availability of Ti). It would seem that in a statistical sense availability of Ti, and the temperature of crystallization are not wholly independent variables, and hence that Buddington, Fahey, and Vlisidis' (1955) generalizations are broadly true, but not necessarily true. In consequence, the magnetite of Tory Island may not have crystallized at the low temperature implied by Buddington, Fahey, and Vlisidis on the basis of TiO_2 content.

EMPLACEMENT OF THE TUFFISITE

Neither the leucocratic nor the magnetitic aplites of Tory Island are ordinary aplite veins. The large proportion of xenolithic and xenocrystic material is very unusual. The mixture of quartzitic and granitic xenoliths is inconsistent with a replacement (granitization in place) origin for the dykes; at least some of these enclosures are exotic. Also, the delicate ramifications of the finest thread-like veinlets are incompatible with magmatic intrusion.

Fissuring and embayment like that illustrated in figure 7b strongly suggests shattering of the rock by explosive gas action, and the sweeping away and recrystallization of the finely comminuted debris. Sometimes plagioclase xenocrysts contain dendritic 'channelways' along which there are no decomposition products, and the feldspar has apparently completely recrystallized (fig. 4b). This suggests intimate penetration along minute fractures.

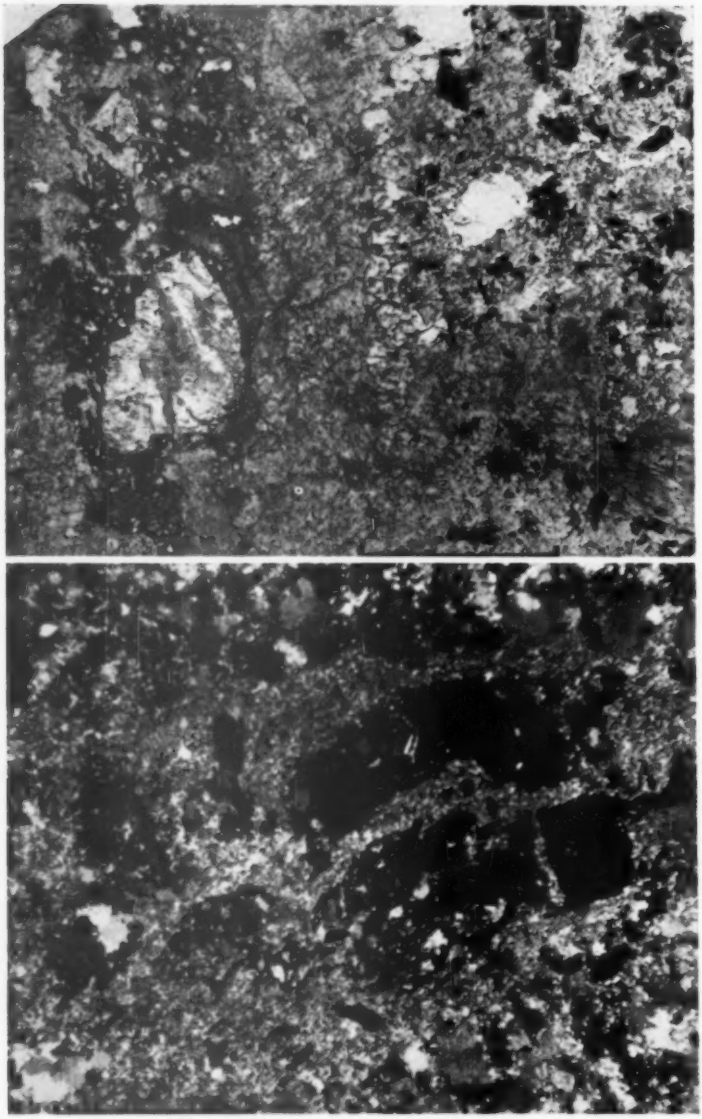


Fig. 7. Magnetite tuffisite specimen 906.

A (above). Photograph (ordinary light) of thin section of a narrow magnetite tuffisite aplitic vein enclosed in shattered recrystallized and aplite-veined granite. Note the microcline xenocryst traversed by a thin tuffisite veinlet. Scale = 0.3 inches.

B (below). Photomicrograph (crossed polars) of part of the field shown in fig. 7A. Fine grained tuffisite veinlets ramify through the large microcline microperthite xenocryst. Scale = 0.18 inches.

Microscopic evidence presented above suggests that the magnetite grew after the main aplite had been emplaced, crystallized, and solidified. Magnetite grew interstitially with a high form-energy, developing euhedra at the expense of the quartzo-feldspathic aplitic minerals. It may be assumed that the magnetite crystallized at a temperature similar to that at which the aplitic material was introduced into the fissures, since the later veinlets freely caused recrystallization and redistribution of the ore. These relationships suggest that the magnetite crystallized from either a gas phase or a very tenuous late-stage magma.

Thus, gas action is thought to have developed the fissures and also produced and introduced the components of the aplites. The intruded material bears a close analogy to ignimbritic extruded material, and is typical tuffisite (Cloos, 1941).

It is difficult to imagine the condition of the matrix during emplacement of the dykes; it was probably largely derived from xenolithic material more finely comminuted than the xenocrysts. Some of the textures simulate those which Reynolds (1951) claimed to be reminiscent of glass shards (see page 121). If these analogies are correct, the veinlets comprised (at least in part) fluid droplets of melted rock which froze during transport and intrusion. Such veins are tuffisites formed by fluidization, in the terminology of Reynolds (1954).

ORIGINS OF MAGNETITE CONCENTRATIONS IN SOME OTHER AREAS

Magnetite concentrations are unknown in other parts of northwest Donegal, except for an erratic boulder of pegmatite containing large quantities of magnetite which Kilroe found at Carbane Hill (see Hull, et al., 1891, p. 109).

Outside Ireland the rocks having closest resemblance to the magnetite tuffisites of Tory Island are the pseudotachylytes described from Parijs by Shand (1916), and subsequently shown by Hall and Molengraaff (1925) to have widespread occurrence throughout Vredefort Mountain Land. Hall and Molengraaff suggested that these rocks are probably most common in marginal portions of the Vredefort granite and in the meta-sedimentary envelope near the contact. The veins contain up to 80 percent xenolithic and xenocrystic material cemented by pseudotachylyte; the larger fragments are as rounded as boulders in a conglomerate. The xenocrysts are salic minerals (quartz, feldspars, or nepheline) taken up from the surrounding country rocks, whilst femic minerals are rare in the veins. The black color of the pseudotachylyte results from multitudes of magnetite granules. Shand noted that the granite walls are not affected more than 1-2 mm from the veins; in some cases the junction cuts absolutely sharply across the mineral fabric of the granite. The pseudotachylyte originated from the granite by melting caused not by shearing but by shock, or alternatively, by gas-fluxing (Shand, 1916, p. 219). Because some xenocrystic plagioclase was melted, and magnetite was derived from biotite by thermal decay, the gas-fluxing must have been at a high temperature. Shand also thought that the spherulitic and microlitic structures, comparable with those in vitreous dyke rocks and lavas subjected to devitrification, imply very great heat during the paragenesis. Hall and Molengraaff's

(1925) alternative explanation of these rocks as flinty-crush rocks due to regional crushing does not match the excellent descriptions and illustrations they and Shand (1916) provide.

From Miller's (1919, p. 518) description, one of the veins in the Lyon Mountain granite, New York, would seem to have certain similarities with the Tory magnetite aplite; this dyke is one of a series of ore bodies within which magnetite is apparently a late crystallization. The dyke is two inches wide, has very fine-grained chilled borders, and a fine- to medium-grained interior. Many sharply defined elliptical to lens-like masses of pink potash feldspar (<¾ inches) occur within the dyke roughly parallel to its walls. Miller believed the magnetite was derived from basic rocks by breakdown of hornblende and hypersthene to diallage, the released iron being taken into solution and introduced at a late stage into pegmatite zones. Gallagher (1937) severely criticizes this view, and claims that there is ample evidence to establish that the principal mineralization resulted from pneumatolytic metasomatism.

Hagner and Collins (1955) considered that Miller's hypothesis " . . . may have been justified . . . ", and thought that at Scott Mine, New York, " . . . the iron in mafic minerals was expelled and became fixed as replacement magnetite. Such a process, controlled by structural features, . . . formed the concentrations of magnetite. During metamorphism and metasomatism, differential movement along planar and linear structures produced high and low pressure zones, and iron migrated to the low pressure zones. Plagioclase is the principal mineral that is replaced by magnetite, and each marked increase in magnetite percentage is accompanied by a comparable decrease in plagioclase percentage."

Besides Gallagher (1937), numerous other authors thought magnetite vein deposits associated with granite contacts resulted from late stage gas action. For example, Wells (1938) claimed that during intrusion of monzonite porphyry in Utah, the hood and overlying rocks of the stocks were fractured, causing sudden release of pressure and liberation of $FeCl_3$ and H_2O gas from the stocks which were still at a high temperature. These escaping gases did not appreciably heat the country rocks, but, among other things, deposited magnetite and haematite veins in the monzonite, and magnetite-haematite replacement bodies in adjacent limestones. Tilley (1951), in describing magnetite deposits at a Tertiary granite contact in Skye, stated the ore skarns " . . . must be attributed to ore-bearing solutions, not directly local in origin, probably ascending from below and attacking preferentially the dolomite which has acted as an absorption apparatus." (p. 647). Reeves and Kral (1955) suggested that tension fractures formed in the cooling outer shell of a Nevadan diorite intrusion, while the central residual magma was enriched in volatiles such as chlorine, fluorine, CO_2, and water, as well as compounds of Na, Fe, P, and Ti. At a late phase the volatiles escaped through the fractured hood of the stock causing scapolitization and deposition of magnetite.

Arriving at analogous conclusions about the magnetite at Granite Mountain, Utah, Mackin (1954) claimed definite evidence for the source of the iron. Mackin said the part of " . . . the mush from which iron was released is now a distinctive rock, light in color and resistant to erosion, that occurs along the

margins of tension joints cutting the interior quartz monzonite. Chemical analyses indicate that this rock contains about 30 percent less iron than the interior quartz monzonite and the peripheral shell; it is the direct source of the mineralizing emanations of the district. A considerable amount of the iron that escaped was deposited at once as crusts of crystalline magnetite on the walls of the tension joints. Most of the veinlets formed in this manner are only a small fraction of an inch in width; . . . the tension joints served as a widely ramifying system of roots that drew from the consolidating quartz monzonite the iron-rich solutions which formed large replacement ore bodies in the limestone at the contact."

These and similar views, based on late-stage ore-bearing solutions (liquid or gaseous), derive considerable support from work on the recent fumaroles in the Valley of Ten Thousand Smokes, Alaska. Magnetite lining fissure vent roofs could not have derived from the nearby pumice, but was obtained from depth by vapor phase activity (Zies, 1924; 1929). Similarly, residual iron and soda rich solutions rising through a dyke-like neck near Klamath River, California, produced magnetitic veinlets (Williams, 1949, p. 29).

In the volcanic caldera of Slieve Gullion, Foughill, and Carrickcarnan (Ireland), Reynolds (1951, p. 132) claimed " . . . the rising gas-streams enriched the Caledonian granodiorite, or glassy tuffisite derived from it, in Si and K, whilst at the same time small quantities of Al, Fe, Ca, Mg, Na, Ti, P and Mn were removed . . . The small amounts of "basic front" materials lost from these low levels, were transported to higher parts of the volcanic superstructure, where they were fixed within basaltic rocks, *e.g.* the olivine-gabbro of Carrickcarnan. These frontal materials, deposited at high levels, can be matched with those found in recent volcanic sublimates, *e.g.* at Vesuvius (Lacroix, 1907), and in the Valley of Ten Thousand Smokes (Zies, 1924, 1929)."

Thus, although it is quite a common phenomenon to find magnetite enrichment near granite contacts, the Tory magnetite vein is unique in Donegal. Hence, widespread concentration of iron-rich fluids or gases capable of reaching the higher levels of the Donegal granites through fractures is unlikely. The only evidence for the source of the iron is seen in the decomposition of granitic mafic minerals (see page 121); this is paralleled by descriptions and illustrations given by Shand (1916), Hall and Molengraaff (1925), and Reynolds

TABLE 3

Grab Samples of Veins of Iron Ore from Bull Valley District, Utah

Specimen	A	B		A	B
FeO	4.55	10.76	Magnetite	14.7	34.6
Fe_2O_3	88.73	73.29	Haematite	78.6	49.4
SiO_2	1.50	12.92	Sphene	0.41	trace
Al_2O_3	1.17	1.27	Apatite	2.7	trace
MgO	0.81	0.22	Pyrite	trace	trace
CaO	1.49	0.03	Quartz	—	11.5
TiO_2	0.17	0.08	Clay	3.0	3.2
MnO	trace	trace	Water	0.33	—
P_2O_5	1.22	0.99			
S	0.03	0.04			
H_2O	0.75	0.50			

(1950; 1951). It may be concluded that production of magnetite resulted from local thermal treatment of mafic minerals during gas action associated with fluidization and tuffisitization.

AGMATITES AND INTRUSION BRECCIAS OF OTHER AREAS

We may follow Wegmann (1938, p. 40) and give agmatite " . . . a merely descriptive meaning, that is to say, to let it denote merely a fragmental rock with a more or less granitic cement, saying nothing about its genesis. In so doing, we may reserve the genetically influenced conception "intrusive breccia" for such cases in which the intrusion mechanics can be ascertained."

Agmatitic rocks are not uncommon in northwest Donegal. Some of these are intrusion breccias resulting from emplacement of the 'Older Granite' (e.g. southeast Tory Island and Rinardalliff Point (Bloody Foreland), and the Tievealehid contact, described by Whitten, 1957a). Others, like the Tory tuffisites, are younger than the crystallization of the granite; good examples occur on the mainland south-southeast of Tory Island.

At Derryconor (Grid reference on fig. 1—165227) two agmatite outcrops occur in the granite adjacent to the granite-quartzite contact. Close to these agmatites the granite and quartzite country rocks are not apparently modified. The breccia comprises a small quantity of interstitial leucocratic granitic material separating angular fragments of quartzite (usually <1 foot in diameter), and occasional small fragments of more micaceous quartzite and semi-pelite. In the tidal reefs to the north very complex relationships exist with confused pelites, scapolitic marbles, appinites (Whitten, 1955), and agmatitic intrusion breccias. At Maley's Rocks (160241, fig. 1) just west of Magheraroarty boatslip, a magnificent intruded breccia occurs. The breccia comprises slightly contorted permeated pelitic rocks incorporated in a very small proportion of fine-grained acid matrix.

Several other examples associated with the Donegal granites to the south have recently been described. Gindy (1951) recorded a sill charged with quartzitic, and various more basic xenoliths at Tallabrista. The Maley's Rocks agmatites are reminiscent of the appinitic intrusion breccia at Kilkenny (Pitcher and Read, 1952). Farther south six agglomerate vents occur roughly along a straight line passing through the center of the Caledonian Barnesmore granite (Walker and Leedal, 1954). These small and irregular vents are enclosed by badly fractured granite, and comprise angular Dalradian schist and amphibolite fragments (<4 feet long, average 6 inches), embedded in comminuted granite and schist. Numerous dykes, bosses, and vein networks, ranging from acid to basic, cut the agmatites. Walker and Leedal thought that wedges of Dalradian schist fragments (which rose at least 1000 feet) were violently driven up into vents within the granite, which probably did not reach the surface.

Transgressive tuffisite veins and dykes have not often been described, although several agmatitic dykes and sills figure in the literature. Although comprising varied components, many of these intrusions appear to have common features; it is proposed to briefly review some recent descriptions.

A number of highly xenolithic and xenocrystic dyke-like intrusions cutting Precambrian rocks in Scotland have been described. One of the earliest references was to a sill at Glen Suileag (Loch Eil district) by Wilson (1898, p. 66). On the west of Glen Creran a small lamprophyric dyke is so full of xenocrysts and xenoliths that it locally resembles a breccia (Muff, 1908). The rounded and angular enclosures were torn from the fissure walls, and carried by the magma towards the upward limit of the intrusion.

Several intrusion breccias are associated with plutonic intrusions in Colonsay. Many of the meta-sedimentary fragments are exotic, and Wright and Bailey (1911, p. 35) suggested that they fell from the sides of a vent blasted out by steam, whereas Reynolds (1936, p. 371) refers to evidence of the enormous upward pressure exerted by the rising magma. Some of the Colonsay lamprophyric dykes contain fragments of quartzite exotic to the island, but Wright and Bailey (1911) were unable to find any evidence to show " . . . whether these fragments have been carried forward from the parent magma mass, or picked up incidentally from Torridonian conglomerates, or derived in some other manner."

From amongst others, three similar examples may be cited. The minor intrusions of Loch Lomond (Scotland) are often densely crowded with cognate and both local and exotic accidental xenoliths, which the magma carried up from very great depths according to Anderson and Tyrrell (1936). Wegmann (1938, p. 86 et seq.) has described basic dykes (<1 meter wide) from Midtluvfik island (S. Greenland) filled with large gneiss and granite fragments which probably originated >1000 meters below the surface. The wider 'brown diabase' dykes of this area commonly caused some melting of adjacent rocks; in large dykes, projecting corners of country rock and local roof zones are often strongly affected. The melted material occasionally occurs as very thin veins in the country rocks, but more commonly produces veins and small intrusion breccias in the diabase (Wegmann, 1938, fig. 47; c.f. Elwell, 1958). Walker and Ross (1954) also noted fusion in a Permian monchiquite dyke near Glenfinnan (Scotland). Xenoliths, ranging from a fraction of an inch to several inches, comprise 50 percent of the central portion of the dyke. Felsitic veinlets of fused country rock ramify xenoliths of Moine granulite. Adjacent to these veinlets plagioclases are corroded and their margins are clouded by partial fusion to colourless glass. Biotite in these xenoliths has been decomposed to an amorphous mass associated with iron ore and ? prehnite.

'Pebble dikes' possess many analogies with the xenolithic lamprophyres referred to above. Farmin (1934, p. 360) defined a 'pebble dike' as a vein-like mass composed of well-rounded to sub-angular pebbles of materials corresponding to formations of the district, several of which are usually represented, in a matrix that is generally a carbonate mud of finely broken limestone fragments, but that also may be a jasperoid gangue or an intrusive igneous rock. In the Tintic district of Utah fragments have been broken from " . . . underlying rocks by magma, or by fluids ejected from the magma during volcanism, and were forced upwards into the rocks in which they are now found in a relatively thick mud . . . In a few of the 'dikes' the matrix surrounding the pebbles is porphyry and the 'pebble dikes' of this type are clear-

ly of intrusive origin, in which pebbles of brecciated country rock were carried along by the advancing front of the intruding magma." (p. 364).

Lovering, Sokoloff, and Morris (1948) and Lovering (1949) referred to the pebble dikes of East Tintic, Utah. The dominant quartzitic pebbles come from the basement several thousand feet below, the larger fragments being ½-10 inches across, but those over 4 inches are uncommon. Some dikes apparently rest on top of monzonite dykes, others are alongside them, and some form a selvedge at the upper or lower side of monzonitic sills. Many have no visible connexion with igneous rocks and some seem to be completely isolated lenses. Lovering (1949, p. 12) believed many pebble dikes " . . . represent material riding on top of monzonite or dragged along the edge of viscous monzonite bodies, but some pebble dikes are probably explosion breccias. It is believed that those of the latter type have formed where superheated steam under high pressure has penetrated into the walls of a fissure that subsequently opened suddenly, allowing an explosive release of the steam contained in the cracks and crevices, thus tearing fragments loose from the walls and injecting them upwards with great violence." Like Farmin (1934), Lovering (1949, p. 13) believed pebble dikes represent a late stage in pre-ore magmatic events.

Tweto (1951, p. 528) in describing a sill of Eagle River porphyry in Colorado, noted an intrusion breccia containing a variety of exotic rocks including well-rounded boulders of chilled porphyry, and pebbles and boulders of strongly altered Precambrian rocks. "The breccia, the presence of Precambrian rocks at this horizon several hundred feet above the basement rocks, and the cross-cutting relations of the ends of the sill all suggest that this sill was emplaced explosively."

There are many analogies between pebble dikes and sandstone (and other sedimentary) dykes, and it would be interesting to know whether there is usually any connexion between the latter and igneous intrusion. Fairbairn and Robson (1942) in discussing the Sudbury (Ontario) breccias claimed that aqueous fluids at elevated temperatures and pressures tore off and comminuted the country rocks to produce a mobile matrix which facilitated transport of the product. Walton and O'Sullivan (1950) recorded a 3-4 inch clastic dyke extending 30 feet up into a dolerite sill. The dyke is continuous with the underlying conglomerate, and it was intruded largely in the solid state while the dolerite was hot and unjointed. Sudden, if not explosive, expansion of water vapor or liquid water in the conglomerate underlying the dolerite momentarily brought solid material in the conglomerate into suspension. This suspensoid swept into the fracture at nearly 575°C and at up to 400 bars to form the dyke.

Walton and O'Sullivan's (1950) example is one of the few sedimentary dykes clearly linked to igneous activity. The connexion may be significant, as in the case of the Utah 'pebble dikes'. A sandstone dyke from West Lothian (Scotland) occupies a fissure which is parallel to a number of nearby igneous dykes; Greensmith (1957) gave a petrographic description of the dyke, but did not record this parallelism. Again, at Soldier's Point at the entrance to Carlingford Lough, Ireland, an undescribed highly calcareous sedimentary dyke traverses Carboniferous Limestone, and is parallel to a Tertiary basic dyke occurring a few yards to the southwest.

Following Cloos' (1941) work on the Swabian tuff pipes, Reynolds (e.g. 1954, 1956) has particularly investigated gas action and intrusion in relation to calderas and central volcanoes. In an actualistic description of Slieve Gullion, Reynolds (1951) described how rising hot gas streams carried detached particles into the smallest fissures. Reference has already been made to petrographic resemblance of the Tory aplites to some of the Slieve Gullion granophyres. From a detailed study of partially fused plagioclases at Slieve Gullion, Reynolds (1952) concluded that the gas temperature was near, and not likely to be lower than, 1355°C (at a pressure of about 300 bars in the presence of water). This remarkably high temperature is consistent with the inference that the eruptions were *nuées ardentes*, and proves that the gas was not derived from crystallizing sub-volcanic magma.

Recently Brindley (1957) described some Caledonian granitic explosion breccias from near Dublin, as the product of an explosive gas phase during the final cooling of the Leinster granite. Analogous with features of the Tory tuffisites are the narrow irregular veinlets of micro-breccia which locally traverse the larger blocks of granite.

Thus, explosive gaseous action apparently produces intrusive xenocrystic and xenolithic rocks under various dissimilar conditions in different areas, and the products may range from rocks of sedimentary to those of igneous appearance. However, there are numerous precedents for invoking gas action for the transport of exotic material from great depth. The breccias of Derryconor and Maley's Rocks (Donegal) are more akin to those described by Farmin (1934), Lovering (1949), Tweto (1951), and Pitcher and Read (1952), but fluidization resulting from gas action was the motivating force. However, the Tory aplitic tuffisites closely resemble certain granophyric aplites of the Slieve Gullion complex, and have less direct analogies with pebble dikes.

CONCLUSIONS

At Tory Island magnetite occurs in veins which originated by forceful gaseous intrusion or fluidization (c.f. Reynolds, 1954). The magnetite crystallized in the interval between intrusion of two successive aplitic phases, and it seems probable that the magnetite resulted from solutions, which may have been gaseous, penetrating the intercrystalline spaces of the first intrusion. Although the Ti-content of the magnetite might suggest low temperature formation, the gases causing the fluidization were almost certainly very hot. It could be argued that such small aplites have a rapid heat loss, enabling the magnetite to have crystallized at a low temperature, but it is more likely that the magnetite crystallized at a higher temperature than Buddington, Fahey, and Vlisidis' (1955) results imply.

Reynolds (1952; 1953) claimed that the Slieve Gullion tuffisites formed at 1355°C and 300 bars; this high temperature has been challenged by Bailey and McCallien (1956). The whole concept of the veins being tuffisites has been discussed, contested, and defended in a series of recent papers (Bailey and McCallien, 1956; Elwell, 1958; McIntyre and Reynolds, 1947; Reynolds, 1937, 1951, 1952, 1953, 1956; Richey, 1937; and Wager and Bailey, 1953). The literature and an admittedly brief field examination causes the present author

to accept Reynolds' conclusions as a working hypothesis. The analogous features of the Tory tuffisites suggest they formed at comparable high temperatures. However, the Tory amphibole xenocrysts are mantled by diablastic amphibole granules (not pyroxene granules, as recorded by Reynolds at Slieve Gullion), which suggests that they were produced at a slightly lower temperature, and in a more hydrous environment, than the tuffisites of the Slieve Gullion caldera. The close analogy with the Parijs pseudotachylytes (Shand, 1916) also supports the claimed very high temperature formation of the Tory veins.

It would be valuable to make a detailed investigation of the physical-chemical conditions under which tuffisites formed. Kennedy's (1948) work on equilibrium between volatiles and iron oxides should be applicable to tuffisites and allied rocks. He suggested that the ratio of ferrous to ferric iron in quenched lavas permits calculation of the partial pressure of O_2 under which the melt crystallized, provided the crystallization temperature can be reasonably estimated. Such an O_2 pressure must be in equilibrium with the remaining volatiles dissolved in the magma, and Kennedy suggests that information concerning gas composition in equilibrium with the lava shortly before solidification can be deduced.

Neither the ultimate nor the proximate origin of the iron is at all obvious. Regional study of the Donegal Granite and its envelope (Whitten, 1957b) demonstrates that there is no correlation between basic rocks enclosed in the granite and the magnetite (c.f. Miller, 1919). Likewise, there is no evidence at the level of exposure to suggest operation of the essentially metamorphic processes postulated by Backlund (1950) and Hagner and Collins (1955). Fracture of the peripheral shell of magmatic granite allowing liberation of concentrated residual iron-rich gases might seem a reasonable explanation of the Tory veins (c.f. Gallagher, 1937; Wells, 1938; Tilley, 1951; Mackin, 1954; and Reeves and Kral, 1955). However, the petrogenesis of the granitic rocks was a protracted process (Whitten, 1957a; 1957b). At an early stage the granite of Tory behaved like a mobile magma, but subsequently it was metasomatized and recrystallized, which, among other things, caused neocrystallization of microcline microperthite porphyroblasts. The aplite veins post-date these porphyroblasts, and it is improbable that at this stage a magma cauldron existed and supplied iron-rich hot gases to one isolated locality. The decay of biotite in the granite adjacent to the veins, and the breakdown of hornblende to aggregates of tiny almost colorless amphiboles, suggest that iron was released and made available for magnetite growth in the tuffisites. By some unknown mechanism magnetite was concentrated in one tuffisite dyke, whilst dispelled from adjacent veins during the fluidization.

The recrystallization and metasomatism of the granitic rocks which are so pronounced around Gola Island (fig. 1), and progressively less well-developed radially outwards, are early processes in the development of material which, by rheomorphism, became potassic granite magma (Whitten, 1957b). The Rosses granitic ring complex (see fig. 1) probably represents regenerated and intruded material of this type. There is no direct evidence of surface expression in the Rosses complex, though it is not unlikely that the complex is

the root zone of a volcano, as implied by Wilson (in Pitcher, 1953b, p. 180).

A profuse swarm of very varied acid dykes is associated with the Rosses complex; " . . . xenoliths are not common though occasionally trains of angular xenoliths are parallel to the walls" (Pitcher, 1953b, p. 167). Acid dykes are also very common on Tory Island and on the north mainland coast to the south. With reference to some arfvedsonitic rhyolite dykes near Tievealehid (fig. 1), Whitten (1954) noted the close association which frequently exists between strongly sodic (particularly riebeckitic) hypabyssal rocks and volcanic ring complexes. It would seem reasonable to link all these acid dykes to the Rosses and Gola complexes, and to claim that these complexes are roots of structures which had surface expression as volcanoes. The Tory Island aplites would seem to be related to this sequence, and to be tuffisites produced from the country rock granite and some Dalradian quartzite.

Although these tuffisitic aplites and Tertiary basalt dykes are closely associated in space on Morard, there is no direct evidence of age in either case; the tuffisites are probably a late stage expression of the Caledonian Newer Granite (of Scotland) cycle. At the same time, the close analogies with the Tertiary granophyres of Slieve Gullion are great, and it must be recalled that Tertiary acid and basic dykes traverse the Barnesmore granite of south Donegal (Walker and Leedal, 1954).

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A PETROFABRIC METHOD OF FOLD ANALYSIS

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ABSTRACT. A petrofabric technique of fold analysis, whereby a quartz fold is "unrolled", as applied to a simple fold by Ladurner (1954), is given in detail. The technique is applied to a more complex fold from the Dalradian of W. Perthshire, in which the degree of preferred orientation of quartz c-axes is weaker than in the fold cited by Ladurner (op. cit). It is concluded that the technique gives results allowing a clear cut interpretation only when one strong maximum occurs in the quartz diagrams.

INTRODUCTION

The detailed method of petrofabric analysis which enables a simple fold to be unfolded was described by Ladurner (1954). Previously, brief reference to the method had been made by Sander (1930, 1950), Knopf and Ingerson (1938), Turner (1948) and Fairbairn (1949).

Recent work carried out in the Petrofabric School at Innsbruck under the direction of Professor Ladurner has shown the limitations of this method when applied to more complicated folds. Folds with a complicated tectonic history from the Dalradian of western Perthshire have been unfolded, but the results obtained do not allow such a clear-cut interpretation as those cited by Ladurner (1954). In some cases however, the results of such analysis are too complicated to be interpreted in the light of present day knowledge.

THE METHOD OF ANALYSIS

The statistical measurement of the fabric of the fold is carried out in the following way. The fold is arbitrarily sub-divided into various sectors (fig. 1a). These sectors should be as small as possible, limited to the smallest area in which approximately two hundred quartz grains can be measured. If the fold is small enough to be contained in one thin section, the sectors are marked off. If they are larger, several thin sections are required. Experience suggests that the fabric of the straight flanks of a fold is homogeneous and only one diagram need be made in these areas. However, the curved axial area of a knee of a fold should be as finely sub-divided as possible. The fabric diagrams of these sectors are then related to the tangent (t) to the fold in the respective sectors (fig. 1a). This tangent can in certain cases be determined accurately by measuring fifty to a hundred micas in each sector and in others it can be determined by megascopic measurement. When the mica diagram is made, the tangent is the trace of the "a b" plane in the projection, i.e. at 90° to the main mica maximum.

Collective diagrams of the quartz and mica measured are then made of the fold in the present form and in the straightened-out state. This latter diagram is made by rotation and superposition of tangents until they lie in a common line, representing the straightened-out state. To illustrate the result on unfolding the fold, a straight line is drawn through the main maxima of each diagram and the centre points of the projection. This line has been called the *q-Richtung* (q-direction) by Ladurner (1954) and is shown in figure 1a by a

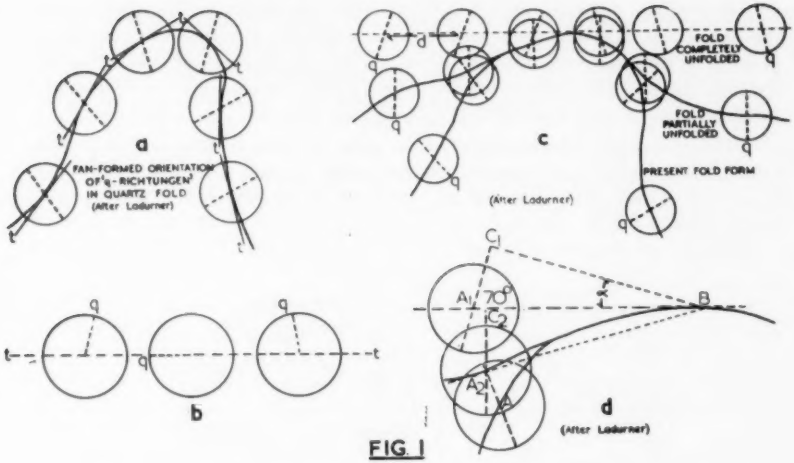

FIG. 1

broken line. In the case of a *Gleitbrett* or shear fold, all the quartz diagrams will show one main set of maxima in the same relative positions. When the fold is straightened out the picture is ideally observed as in figure 1b. In an ideal flexure fold, however, the main maxima are approximately co-linear when the fold is unfolded.

A quartz fold studied by Ladurner (1954) was divided into six sectors and the *q-Richtungen*, when these sectors were in their original positions, were approximately fan-shaped in distribution (fig. 1a). When the fold was completely unfolded no correspondence existed in the position of the correlated maxima on the common tangent (t) but the individual *q-Richtungen* lie symmetrical to the top of the fold. In this instance the angle which these *q-Richtungen* make with the tangent (t) is extremely constant *viz*: 73°, 70°, 73°, 70°, 68° and 80°; such folds have been termed *Fächerfalten* (Sander, 1930) or fan folds. If the fold is only partially unfolded, that is until the correlated *q-Richtungen* are aligned parallel to one another, a shallow troughed fold-form is obtained (fig. 1c). It is now required to bring the *q-Richtungen* parallel to one another (in this case normal to A_1B, fig. 1d) in such a way that the distance between the centre points of individual diagrams remains constant, i.e. equal to 'd' (fig. 1c). This is done by constructing a right-angled triangle with A_1B as the hypotenuse, so that one side lies parallel to the *q-Richtung* through A_1 and forms a right angle with C_1B at C_1. Then, with the top of the fold B as rotation point, the triangle A_1C_1B is rotated until the line A_1C_1, which includes the *q-Richtung*, lies perpendicular to A_1B. The *q-Richtung* now lies in line C_2A_2 and line A_2B is tangent to the partially unfolded fold at the point A_2. A, A_2 and A_1 are therefore the middle points of the sectoral diagrams of the present fold A, the partially unfolded fold A_2 and the completely unfolded fold A_1. The angle α is the complement of the angle between q and t.

This procedure is carried out in the same way for all diagrams of the

sectors and the partially unfolded form is thus constructed. The distances of the individual diagrams from each other are thereby maintained as closely as possible considering the varying curvature of single parts of the fold. This analysis significantly shows that the field is homogeneous only in the partially unfolded state and by further straightening out becomes heterogeneous.

The symmetrical arrangement of the *q-Richtungen*, that is the symmetrical orientation of quartz maxima to the fold knee in both flanks, indicates that these q-directions were already in existence before the bending to form a heterogeneous fabric. The fold thus attained its present form through a period of shearing, giving rise to the form as reconstructed in the partially unfolded

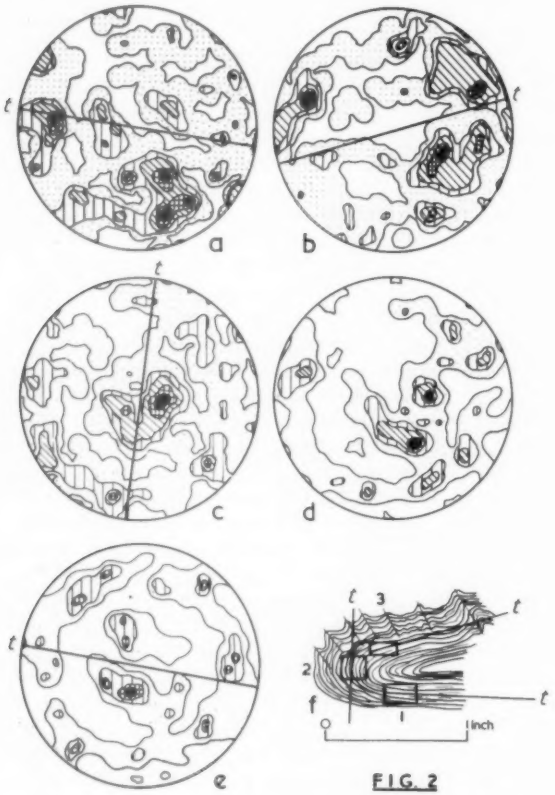

FIG. 2

Fig. 2. a. Quartz; 200 'c' axes. Contours. 3.5, 3.0, 2.0, 1.5, 0.5, 0.0%.
b. Quartz; 200 'c' axes. Contours. 3.5, 3.0, 2.0, 1.5, 0.5, 0.0%.
c. Quartz; 200 'c' axes. Contours. 4.5, 3.5, 2.5, 1.5, 0.5, 0.0%.
d. Quartz; 600 'c' axes. Contours. 2.66, 1.75, 1.66, 1.0, 0.15, 0.0%.
e. Quartz; 600 'c' axes. Contours. 2.3, 1.75, 1.66, 1.0, 0.15, 0.0%.

state, followed by flexure folding with the retention of the old fabric axis and the "a c" symmetry plane.

This example, used to illustrate the method of analysis, lends itself to a precise interpretation due to its simple tectonic history.

APPLICATION OF METHOD TO A DALRADIAN FOLD

The results obtained by applying this method to a more complicated fold from the Dalradian of western Perthshire are now described and discussed. The fold (fig. 2f) came from Coire Chaorach 1 mile east of Ben More and occurred in the Ben Ledi Grit Group.

The fold was divided into three sectors (1, 2, 3, fig. 2f) since it was too small for further subdivision. Since the fold is not simple, but is complicated by a series of minor folds superimposed on the major one, muscovite diagrams were made to prove statistically the direction of the tangent to the part of the fold measured in the three individual sectors. The muscovite diagrams are simple, showing a clear S tectonite form; the knee of the fold however shows a less perfect form than the other two.

Description.—The quartz grains in the three sectors were measured and the results are now described.

Figure 2a (*Sector 1*). The diagram shows four maxima, three of which have no separate significance and can be grouped together as one maximum group, M III in the classification of Sander (1930). The other maximum is M I (Sander, 1930). This lies in, or close to, the principal S plane in many S and B tectonites. The rock is a B tectonite.

Figure 2b (*Sector 3*). This diagram shows four maxima, two of which belong to one maximum group and the other two of which can be treated separately. The diagram shows a distinct B-tectonite form with a minor concentration around the mica maximum in the corresponding muscovite diagram. M I is again represented, but in this case is part of a larger maximum group, deviating by 25° from the projection of the main S plane.

Figure 2c (*Sector 2*). The fabric shown by this diagram has a much better B tectonite form than those of Figure 2a and 2b. A submaximum M I lies in the main S plane. The main maximum lies in the small circle (*Kleinkreis*—Sander, 1930) 20° from the main S plane, in a plane normal to it. The significance of this small circle is not fully understood. However, it may be the result of a relict sedimentary fabric or a pattern caused by *Schnitteffekt* (Sander, Kastler, Ladurner, 1954). The *Kleinkreis* is much better developed than the main peripheral girdle.

Collective diagrams of quartz and mica were made for the fold in its present form and in the unfolded state. The quartz diagrams are given in figures 2d and 2e respectively. These are now described.

Figure 2d. A poorly developed girdle and a *Kleinkreis* around the B axis are seen in this diagram. Two maxima occur in the girdle of the *Kleinkreis.*

Figure 2e. A well developed girdle occurs around the B axis and an equally well developed *Kleinkreis* is also present. One maxima is shown and this lies approximately in a plane normal to the main S plane. The *Kleinkreis* is divided into two parts by the principal S plane (or the tangent to the fold).

Discussion: The two collective diagrams figures 2d and 2e are not essentially different, but the better development of an "a c" girdle in the unfolded state indicates that an "a c" girdle was in existence before the formation of the fold in its present form. Figure 2d shows that the final flexure movement partially destroyed the well developed girdle previously existing. The B axis, however, remained the same during the episodes of deformation. It is suggested that the two episodes were approximately contemporaneous and that one developed from the other.

The fold can be classified as a heterogeneous flexure fold (*Inhomogen Biegefalte*—Sander, 1930). The fabrics of the individual sectors of the folds are heterotactic as shown by the muscovite and quartz diagrams.

CONCLUSIONS

The fold analysed by Ladurner (1954) can be considered a type example of a fold formed by an initial shearing followed by simple flexuring. The diagrams are simple in form and only one maximum occurs in each of the diagrams made. Thus the *q-Richtung* was established with certainty. The fold from the Dalradian, however, is of a type more often encountered in nature and the *q-Richtung* could not be established with the same precision and certainty, because the individual diagrams show more of a B tectonite form than those examined by Ladurner. Thus the method as applied by Ladurner to a simple case can still be applied to more complicated folds but in a modified form. To unfold a fold of this type the overall patterns of the diagrams from the sectors had to be used. In the Dalradian case described above a clear fabric does not emerge on analysis for a number of reasons, which may, singly or collectively, be effective.

(1) The rocks have mostly been subjected to four episodes of deformation. This has been established by an extensive study of mega-, meso-, and microscopic fabrics of the Ben More area. The relative importance of each episode in the build-up of the present rock fabric is difficult to assess.

(2) The quartz grains in the rocks often show an inequidimensional form and it is difficult to estimate the degree to which the diagrams have been modified by *Schnitteffekt* (Sander, Kastler, Ladurner, 1954; Kaemmel, 1955).

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ON THE TiO_2-CONTENT OF MAGNETITES AS A PETROGENETIC HINT

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ABSTRACT. Examination of the TiO_2-content of magnetites extracted from different rock types of the same schist belt shows that there may be, within the same orogeny, a close relationship between this TiO_2-content and the original material of the host rock.

INTRODUCTION

The titanium-content of magnetites of common rocks may vary within wide limits. Buddington, Fahey and Vlisidis (1955) suggested that these variations may be used as means of interpretation of the temperature of formation of the host rocks. This possibility is very attractive and interesting, and it really exists if the Ti-content of magnetite is higher than that which for a given set of geological conditions remains as solid solution in magnetite, so that in the Fe_3O_4—$FeO.TiO_2$ system, Ti can stay in solid solution only if the conditions of crystallization of magnetite favor that instead of an exsolution of ilmenite.

Heier (1956), however, argued against such a possibility, paying thereby more attention to compositional factors than to the temperature.

The magnetites discussed by Buddington, et al. (1955) are all rich in titanium. Also the magnetites in the granitic rocks they discuss contain 0.9 to 9.6 percent TiO_2, the lowest value reported being that for the magnetite of a microcline-oligoclase alaskite.

In this respect, the magnetites of the Sierra Leonean rocks, examined by the present author, are entirely different. There the TiO_2-content of magnetites analyzed is, in most cases, below 0.5 percent and the highest value obtained was less than 2.0 percent, and still there seems to be a reasonable correlation with certain geological facts.

This finding prompted the writing of this paper.

MATERIAL EXAMINED

The Sula Mountains and the Kangari Hills form, in Central Sierra Leone, a rather extensive schist belt, probably representing a part of an old Precambrian geosynclinal formation. This schist belt is embraced by synkinematic granodiorites, granites, and porphyroblastic varieties of granitic rocks. In addition, there are minor latekinematic dykes and bosses commonly associated with pegmatites of simple type, some of which contain however, disseminated coarse magnetite.

From this area, 22 rock samples were chosen for examination (see the list of table 1). From each sample the magnetite was separated using an electromagnetic separator (mainly by Mr. J. Middleton of Geol. Survey of Sierra Leone), and analyzed for TiO_2 (in 6 cases also for MnO). The concentration analyzed contained only negligible amounts of any silicate material.

Most samples were analyzed colorimetrically by Mrs. Irja I. Huhta and Mr. John Middleton, both of the Geological Survey of Sierra Leone.

RESULTS

The three first-mentioned analyses of table 1 represent rocks of a continuous and more or less coherent serpentinite zone (Marmo, 1958a). Also the

TABLE 1

TiO_2 and MnO Content of Magnetites of the Rocks of Central Sierra Leone, and Fe_2O_3 and TiO_2 Content of the Respective Rocks

Host rock	In magnetite		In the host rock	
	TiO_2	MnO	Fe_2O_3	TiO_2
1. Olivine-tremolite serpentinite. A good representative of the Kangari Hills serpentinites, which form a continuous strip several km in length and 3 km in width. (Marmo, 1958a)	% 0.24	%	% 5.48	% 0.16
2. Tremolite serpentinite of the same strip as in (1)	0.20		3.80	0.08
3. Talc-anthophyllite rock, asbestiferous. From a strip, 2 km long and 0.2 km wide, adjoining the serpentinite belt of the Kangari Hills (Marmo 1957)	0.24			
4. Magnetite-chlorite schist from a shearzone adjacent to serpentinites. Yirisen	0.86			
5. Fine-grained amphibolite showing pillow-lava structure. Interior of the Kangari Hills. Ancient basaltic flow	0.48		4.58	0.79
6. Cordierite-andalusite schist. Spotted schist ("Knoten schiefer") enclosed by granodiorite, at the southern end of the Kangari Hills	0.41			
7. Garnet-cummingtonite-magnetite schist. The Kangari Hills	0.27	0.21	9.06	0.30
8. Banded ironstone adjoining (7)	0.24	0.12		
9. Diorite. Probably intrusive. Makeni, resthouse	1.97	0.40	1.90	0.72
10. Diorite gneiss. Metasomatic. Near Bumbuna	0.19			
11. Synkinematic biotite granite. Gorahun. Near diorite of Jagbura, which is probably intrusive	1.17		0.71	0.19
12. Synkinematic biotite-muscovite granite. Metasomatic. N. of Masimo	0.29		1.75	0.05
13. Synkinematic biotite-muscovite granodiorite. Metasomatic. Kurumanto	0.26		1.06	<0.01
14. Porphyroblastic granite. Bahama. Close to (11)	0.77		0.99	0.50
15. Porphyroblastic granite. Bokuma. Within metasomatic rocks	0.17	0.09	1.58	0.22
16. Porphyroblastic granite. Near Kakonta, E of the Seli River	0.20	0.11		
17. Latekinematic granite. Related to helsinkite. Dalakuru	0.42		2.16	0.24
18. Latekinematic albite-epidote granodiorite. Near Kunya	0.40		1.70	0.10
19. Latekinematic granite. From a fault in the Teye River	0.46		0.41	0.08
20. Latekinematic migmatizing granite, cuts pegmatite. Badja	0.49		0.43	0.07
21. Magnetite pegmatite. Bumbuna	0.2	0.2		
22. Magnetite pegmatite. Badja	0.26			

Analysts: 1-3, 17, 22 Mr. John Middleton of Geol. Survey of Sierra Leone
4-12, 14-15, 20 Mrs. Irja I. Huhta of Geol. Survey of Sierra Leone
13, 16, 18-19 The Author
21 Dr. Oleg v. Knorring of the Leeds University

magnetite-chlorite schist (Anal. 4), which contains magnetite octahedra up to 0.6 cm across, belongs to a sheared portion of the same serpentinite belt. The talc-anthophyllite rock (Anal. 3) is situated at the verge of the serpentinite belt at a place where tension prevailed during its formation, anthophyllite probably being formed from antigorite, and, locally, further altered into talc (Marmo, 1957). Thus it is reasonable to assume that in all these cases, magnetite has been formed under more or less similar conditions and within petrochemically similar environment. For the magnetites of serpentinite and of talc-anthophyllite rock, such a possibility is rather well confirmed by the TiO_2-content of their magnetites, which contain 0.20-0.24 percent TiO_2. Furthermore, in both rocks mentioned, the Ti seems to be entirely bound in magnetite, because no actual Ti-mineral has been detected microscopically. The 3½ times larger TiO_2-content appearing in the magnetite of the magnetite-chlorite schist is, from this point of view surprising, and it will be discussed later.

The fine-grained amphibolite of sample 5 is interesting, because this rock is of undoubtedly volcanic origin (pillow-lava beds).

The similarity of the magnetites of Anal. 7 and 8 is expected because, in Central Sierra Leone, the banded ironstones are actually parts of garnet-cummingtonite-magnetite schists, representing portions of the schists that are especially enriched in magnetite.

The granites of Central Sierra Leone could be classified into syn- and latekinematic granites (Marmo, 1955) if the classification proposed by Eskola is used. This classification seems to suit the circumstances of Sierra Leone rather well. The synkinematic acid rocks are mostly of granodiorite composition. The potash feldspar of these rocks is microcline and undoubtedly metasomatically introduced. In many cases these rocks could be proved in the field to have originated from sediments. In such cases, there are abundant and well preserved remnants of schists enclosed by granodiorites, into which they texturally grade. Many of the porphyroblastic granites of the same area belong to the same category (Marmo, 1956). The diorite gneiss of Bumbuna is certainly metasomatically formed, and it belongs to the mantle part of a concentric structure resembling a dome. In the synkinematic rocks, the TiO_2-content of magnetites is very variable: 0.17 to 1.17 percent.

Among the synkinematic acid rocks, however, most of them have been derived from sediments, although some may be products of granodioritization or granitization (Marmo, 1956a) of some intrusive rocks. The granitic rock with magnetite containing 1.17 percent TiO_2, for instance, in the field seems to be closely related to a diorite body of intrusive character. A very similar diorite has a magnetite with 1.97 percent TiO_2. The porphyroblastic granite of Anal. 14 (0.77% TiO_2) is a portion of the above-mentioned granodiorite. Thus these analyses, in the opinion of the present writer, rather indicate that the original material of these rocks has been intrusive—an intrusive diorite or the like. The rest of the analyses of the synkinematic rocks, on the other hand, correspond to rocks of sedimentary origin, as they contain comparatively well preserved schist remnants.

Regarding the latekinematic granites (Anal. 17 to 20), they are all geo-

logically very similar rocks, which are pink, medium-grained and tend to be aplitic. Two of them are albite-epidote granodiorites rather than granites, and actually they approach, locally, in respect to their mineralogical composition, the helsinkites. The migmatizing granite, on the contrary, is a real latekinematic microcline granite, also rather aplitic. It appears in the field migmatizing the synkinematic granodiorite and also penetrating magnetite-bearing pegmatite. Magnetite extracted from all these definitely latekinematic grantic rocks is surprisingly similar, insofar as their TiO_2-content is concerned, varying between values of 0.40 and 0.49 percent TiO_2.

Pegmatites are represented by the analyses of 21 and 22 (table 1). They are simple pegmatites mainly consisting of quartz, feldspars, and micas, but containing disseminated coarse magnetite (up to 1 inch across). These samples come from localities several tens of miles apart, on the opposite sides of the schist belt. Bumbuna is approximately in the middle of the western flank of the schist belt, east of the Sula Mountains, whereas Badja is some ten miles east of the Kangari Hills, in an area mainly composed of the synkinematic granodiorites but cut by dykes of latekinematic aplitic granite as well (Anal. 20). Yet the magnetites extracted from these two pegmatites have a very similar TiO_2-content (0.20 and 0.26%).

Cordierite-andalusite schist (Anal. 6) forms strips within granodiorite of sedimentary origin. In places these strips form continuations of the garnet-cummingtonite-magnetite schists, but they also occur close to the fine-grained amphibolites of the southern Kangari. This particular magnetite was extracted from a sample taken from a strip entirely enclosed by granodiorite.

DISCUSSION

Thus among the granodiorites of Central Sierra Leone, some have magnetite of comparatively high TiO_2-content, and some of rather low TiO_2-content. Furthermore, the latter granodiorites and porphyroblastic granites could be interpreted as being of definitely sedimentary origin, but the former ones—those having magnetite rich in TiO_2—are of less certain origin; since they occur with intrusive diorite bodies, they may have originated by granodioritization of intrusive rocks. This assumption is confirmed by the comparatively high TiO_2-content of the magnetite extracted from the above-mentioned intrusive diorite.

In the opinion of the present writer, this would mean that in Central Sierra Leone, there are granodiorites and porphyroblastic granites primarily derived from material of two different kinds, and that these granodiorites and porphyroblastic granites may be distinguished from each other also by the titanium-contents of respective magnetites. A lower TiO_2-content of magnetite would thus indicate that the primary material was sedimentary; a higher content, on the other hand, would mean that the rock is a product of granodioritization or granitization (Marmo, 1956a) of some intrusive material; judging by the TiO_2-content of the magnetite of an intrusive diorite (Anal. 9), this material may have been dioritic or quartz-dioritic.

The latekinematic granites (Anal. 17-20) and pegmatites (Anal. 21 and 22) have a magnetite poor in titanium. This does not necessarily mean that

these rocks were formed at low temperature because it may, and probably does, depend upon the deficiency of titanium in the material which yielded the latekinematic granites and pegmatites. On the other hand, these latekinematic rocks are typical of the microcline granites characteristic of many Precambrian areas, which probably have been emplaced under moderate temperature (Marmo, 1958b). The latekinematic granites considered in this paper may have different mineralogical and petrological composition, but they still behave in the field in exactly similar ways, and the TiO_2-content of magnetites extracted from these rocks is in all cases very much the same.

The two pegmatites here considered are similar in all respects and their magnetites are equally rich in titanium, although they occur on opposite sides of the large schist belt.

Regarding the magnetite of the fine-grained amphibolite of basaltic origin, the somewhat higher TiO_2-content is to be expected, because in extrusive rocks, according to published data, the titanium is present in higher amounts than in other rocks.

The exceptionally high TiO_2-content of the magnetite-chlorite schist is surprising, because geologically and according to the observations in the field, this schist is closely related to the serpentinites containing magnetite of considerably lower TiO_2-content. By analogy with the synkinematic acid rocks, this discrepancy may also depend upon different primary material.

CONCLUSIONS

The examination of the TiO_2-content of magnetites extracted from different rock types belonging to the same schist belt has given the following results, which, however, must be considered as valid only for the rocks of that belt.

1. Rocks of the same type and origin, and belonging to the same orogenic schist belt, have magnetites of equal TiO_2-content.

2. If the TiO_2-content of magnetites extracted from rocks of similar composition and lithology are dissimilar, the rocks may have been derived from different primary material.

3. In the case of the acid synkinematic rocks (granodiorites, granites, and porphyroblastic granites), the different TiO_2-content of respective magnetites may make it possible to distinguish rocks derived from sedimentary material from those derived from primarily igneous material.

4. Composition as well as temperature may determine the composition of magnetite where $FeO.TiO_2$ is inadequate to saturate the magnetite for the temperature and pressure involved.

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ESSAY REVIEWS

The Future Supply of Oil and Gas, a study of the availability of crude oil, natural gas, and natural gas liquids in the United States in the period through 1975; By BRUCE C. NETSCHERT. P. xi, 134; 19 tables. Baltimore, 1958 (The Johns Hopkins Press, for Resources for the Future, Inc., $3.00).—This crisis in the Middle East emphasizes once again the influence of oil in world affairs. Any estimates of future supply are certain to be scrutinized—doubly so if published under reliable auspices and if they differ or seem to differ from previous conclusions.

Bruce C. Netschert's small readable book falls into this category. Specifically, he concludes that by 1975 about six billion barrels of crude oil a year and 22½ trillion cubic feet of natural gas a year could be available for production in the United States, at no appreciable increase in constant-dollar cost. In view of past estimates that peak production is expected to be only about half those amounts, and that the peaks may be reached as early as the 1960's, Mr. Netschert has thus written a controversial book that is sure to be quoted and misquoted for a decade or more.

Mr. Netschert is a research associate on the staff of Resources for the Future, Inc. Both by training and vocation he is a professional minerals economist but has some training in geology. His present book is part of a larger study of energy commodities, being made by the energy and minerals staff of Resources for the Future, in relation to the expected development of the American economy through 1975.

Resources for the Future is a non-profit organization whose purpose is to advance the development and proper use of natural resources, through identification and analysis of the problems involved. This is the aim of the book under review, with regard to oil and gas. The book takes nothing for granted—it goes back to definitions and concepts and attempts to make a clear distinction between fact, inference, assumption, judgment, and guesswork—explaining along the way the technologic and economic milieu in which these are applied.

The theme is the availability of petroleum and natural gas by 1975, as controlled by the natural stocks of oil and gas in the ground and by the technology of finding and producing them. Here the author encountered a fact that even geologists and engineers must make a point to remember: To a petroleum man *reserves* mean *proved recoverable reserves*, whereas to others the term is less restrictive. To provide a common base of understanding, Netschert adopts the terms *reserves*, *resource base* and *resources*. Reserves are the stocks of oil and gas producible with current techniques at costs consistent with normal commercial risk-taking. *Resource base* is defined as including all the oil and gas in the earth's crust, while *resources* constitute that part of the resource base (including reserves) likely to become available under given technologic and economic conditions.

A second and more compelling reason for adopting new definitions was to allow for improvements in discovery technology, for

> In the numerous discussions of the probable course of future oil [and gas] discoveries the tendency has been . . . to hold technology as well as costs constant . . .

The author methodically analyzes all recent estimates of future reserves and production, starting with the Weeks estimate of 1948. These he converts

into their "resource-base equivalent" and then considers the likelihood of improvements in discovery and recovery technology in order to reach some estimate as to what part of the resource base may actually be available if there is need to go after it. For oil, he concludes:

> The total crude oil awaiting . . . future recovery in the United States can be inferred . . . to be on the order of 500 billion barrels. This includes present proved reserves, the currently unrecoverable content of known reservoirs, and [200 billion to 300 billion barrels yet to be discovered and discoverable by present technology].

To some observers the fact that oil is more difficult to find now than in the general past is "an indication of scraping the bottom of the barrel". A more precise statement of the case, I think, would be simply that improvements in our skill at discovering oil are not keeping pace with exhaustion of favorable sites. That is, the blame is not all Nature's.

Netschert lists four factors indicating the probability that resource limits are not in fact being approached. "A large portion of the area of the United States (particularly the continental shelf) is geologically favorable . . . but has not been given the intensive exploration and development effort of present and past producing 'provinces'." Secondly, "there has been a failure in the past to explore known structures thoroughly, and . . . a preoccupation with some types of geologic traps at the expense of other types", specifically structural traps as against stratigraphic traps. A third factor is that of depth, "which presents possibilities in both known and unexplored areas", and the final factor, for which he cites Levorsen as his source, is the "oil potential in such currently novel sources as pre-Cambrian and basement rocks, crystalline rocks, volcanics, and continental and freshwater deposits."

Advances in exploration technology can be expected that will tend to balance out increases in exploration costs—nothing spectacular, but "steady refinements of existing techniques, [including] better integration of geology and geophysics". The most important future tool of discovery, however, will continue to be the drill. Improvements in drilling practice, too, are on the way, with consequent benefits in lower costs. The requirement for an ever greater number of wells, drilled deeper and under more difficult conditions (as on the continental shelf), constitutes a persistent pressure to which the industry must respond.

As to recovery:

> A recovery factor of two-thirds as the average for the United States within the next fifteen years or so [as compared with a present average variously estimated between about 33 and 40 percent] would be well within the possibilities currently being opened up.

Netschert concludes finally: (a) that the ground contains enough oil to meet any demand through 1975; (b) that improvements in discovery technology will progressively increase our ability to find it; and (c) that with the help of improvements also in recovery technology the United States should be able, if needed, to produce about six billion barrels of crude oil a year by 1975, at no appreciable increase in cost except as may come about by inflation. This figure of six billion barrels a year is composed of four billion barrels of primary production and two billion barrels of secondary production.

These figures compare with the published forecasts that actual United

States production will reach a peak variously estimated at 2.0 billion to 4.1 billion barrels a year, the peak year coming as early as 1960 according to some, and possibly as late as 1980 according to others.

A great deal of judgment is involved in this verdict—and also a great deal of subjectiveness. Although Netschert goes through a lengthy process of trying to develop a figure that would describe the basic stock of oil available for production, he draws no inference as to how much production such a base could support; instead, he chooses to select the 1957 production estimates of A. Cadle and D. C. Ion (which happen to be the highest of the lot), because they appear to him to constitute a "reasonable quantitative approximation" of his general conclusions, as cited in (a) and (b) above. Moreover, he fails to point out that the Cadle and Ion estimates represent both primary and secondary production, whereas he applies the figure to primary production alone. He admittedly sides with the optimists—his word—in concluding that large additions to production will be possible through improvements in secondary recovery, acknowledging a temerity—again his word—that the experts themselves do not display. He simply has faith in technologic progress. There is nothing wrong in that—many of us have such faith—but it is necessary to note that this faith, rather than fact or judgment, is at the base of this particular conclusion.

Perhaps not even time will tell whether Netschert's faith in technology is soundly based or not. The necessary pressure to produce may or may not spark the explosion that he foresees. Other sources of supply—imports and oil from shale—will have a bearing both on the quantity actually produced, and at what cost.

The situation with respect to gas is even less easy to analyze. One peculiar feature is that predictions of supply are based on some figure connoting the number of cubic feet of gas expected to be produced per barrel of oil produced. None of the statistical gas-oil ratios is wholly satisfactory, mainly because not all gas is associated with oil, and so Netschert makes his own selection—but in applying it to his previous oil estimates he gets an implied gas availability in 1975 that is higher than he is willing to accept; instead he adopts the highest figure published, namely, the estimate of the American Gas Association, 22½ trillion cubic feet.

As with oil, the amount of gas that is in fact produced will depend on the availability of imports and substitutes. "In the coming decades imports from Canada and Mexico should become a formidable competitor . . . in several important regions." There may be imports also from Venezuela and even from the Middle East, transported in liquid form. As to synthetic gas, however, it "is unlikely to provide more than local competition by 1975".

Netschert concludes his book with a discussion of natural gas liquids (N.G.L.), giving them the same coverage—reserves, future supply estimates, technology, and other sources of supply—as was given oil and gas.

Most of the past estimates—forecasts, projections—of future production of oil and gas that Netschert analyzes are based on what he terms the "decline-curve" technique, that is, on the assumption that production over a period of

time will follow a curve that rises to a peak and then declines. Mr. Netschert does not approve of the decline-curve technique. He disapproves of it so thoroughly, in fact, that he spends more than five pages criticizing it, losing in the process the analytical detachment displayed generally in the book. There are two schools of thought about statistical analysis, one composed of those who believe in detailed weighing of each factor, to which school Mr. Netschert belongs, and the other composed of those who believe in projecting gross trends. Expert preference can be found on both sides, and Mr. Netschert's attack on the other side's methodology is unworthy of him.

There is no necessary conflict between Netschert's estimates and the much lower previously published forecasts, although such conflict is certain to be read into them. Netschert's forecasts apply to the amount of oil and gas that can be produced in 1975 if there is the need for it; the others apply to what is actually expected to come out under the supply-demand conditions of the time. Both sets of estimates are mixtures of judgment and bias, in part professional bias and in part strictly personal inclination, as pointed out above.

In fully assessing the book, one must remember that the author's conclusions are backed by an exhaustive search of the literature; in addition, he had the advantage of pre-publication comments by most of the authors whose estimates and predictions he discusses, and by many others as well. Fully 60 separate papers are cited, no mean coverage, considering that the problem is perhaps more technologic than economic and one in which technical men themselves flounder.

The treatment is largely philosophical and analytical, but with enough technology to add spice. The clean, readable descriptions of technology, understandable by any intelligent layman, could only have been written by someone having a grasp of his subject. I encountered only two words that would give a lay reader difficulty. At only a few places is the meaning or logic not immediately clear. Here and there the author argues at cross-purposes with himself, but still less than one often sees in technical papers by authors professionally expert in their subject.

Viewed in the above light, the book is a remarkable achievement. The reader may find fault here and there, but no technical work ever written has satisfied every reader in all its nuances.

The editors and publishers also are to be congratulated. The book is printed in handy, almost pocket, size; arrangement and style are attractive; the type is easy to read; and the tables are well arranged. The over-all impression is so pleasing that one feels a pang of disappointment at noting an occasional imperfection, such as a discrepancy between tables and narrative. Another imperfection is the abundance of topical material in footnote form. To my mind, such material belongs in the body of the text, but here, too, I confess, there are two schools of thought.

In sum: No one who has an interest in the future supply of energy in the United States—whether his interest be practical or academic, whether he be a producer of oil and gas or a consumer, a maker of policy or one to be affected

by it, and whether he agrees with Netschert's conclusions or not—can consider himself fully informed on this subject until he has read Netschert's book. And not only will he find the reading profitable, he should find it enjoyable as well.

SAMUEL G. LASKY

Osteology of the Reptiles; by ALFRED SHERWOOD ROMER. P. xxii, 772; 243 figs. Chicago, 1956 (University Chicago Press, $20.00).—This volume undoubtedly contains the most extensive systematic account of the skeleton of both living and fossil reptiles ever assembled. Both paleontologists and herpetologists will find in it a wealth of information on the structure and variations of the reptilian skeleton. Professor Romer has compiled most of the available data on reptilian osteology and added no little new material, particularly on certain living groups. Following the general plan of the posthumous work by S. W. Williston that bears the same title, the book consists of two parts: a comparative account of the bones of the reptile skeleton, region by region, followed by a systematic review of the Reptila with osteological diagnoses of all taxa down to families and a complete list of genera. Each of these is a major work in itself and Professor Romer is to be complimented for successfully assembling and digesting this vast array of data and presenting it in readable fashion.

The section on comparative osteology opens with a summary of those parts of the soft anatomy of reptiles that may influence the skeleton, followed by some generalities on skeletal tissues and the individuality of bone elements. *Seymouria*, an early Permian fossil whose assignment to the Amphibia or Reptilia is uncertain, is used as the prime example of the basic reptile skeleton; this account is supplemented by description of the skull in the extinct captorhinomorphs, another early and primitive group, and finally by descriptions of living reptile skulls. Although this treatment has the dual advantage of commencing with the most primitive known types of reptile and also the most thoroughly ossified skulls, interpretation of these fossils ultimately rest upon our knowledge of the structures of living forms. A lizard or *Sphenodon* might have been a better point of departure than a fossil in discussing the anatomy of both skull and postcranial skeleton.

Following these detailed descriptions, the skulls of the better known representatives of each order and major suborder of reptiles are figured and their salient features discussed. Use of the Pseudosuchia as a basis for comparison of other archosaur skulls, although logical, is unfortunate because they are so inadequately known. After the skulls, vertebral column, ribs, girdles, limb bones, and feet are treated in similar fashion; the basic pattern in cotylosaurs is described in detail followed by comparative treatment of the same series of types. The chapter on vertebrae is perhaps the least inspired; failure to indicate size of various vertebrae (all drawn to the same height) is in places confusing and apt to be misleading, and more complete labeling of some illustrations to show special terminology of vertebral processes would have been most helpful to the reader.

The chapters on limb girdles and limbs are far more lucid. A condensed but comprehensive and interesting essay on the locomotor mechanisms of tetrapods opens the chapter on limbs. It is followed by a detailed functional

analysis of the limbs which essentially repeats the treatment of this topic in the author's "Vertebrate Body", and seems out of proportion to the space accorded to such consideration of other regions of the skeleton. Of interest here (p. 337) is the suggestion that limb diameter of land vertebrates increases as the 1.5 power of the linear dimensions of the animal. This is far more reasonable than Professor Romer's earlier proposal to relate linear measurements of bones to a unit which was the 2/3 power of average dorsal vertebral diameter. Application of this suggestion is necessarily omitted from the book. Nor does he discuss the possibility that variation in limb proportion may be at least partly a function of differential growth rates; basic observations of growth rates have not yet been made but offer a fruitful field for further investigation. Other features of special interest are the figures of ribs, which suggest that these structures might be of greater value in taxonomy than has generally been thought, and the compilation of numbers of presacral vertebrae in various families.

The critical reader will not some inconsistency between the examples described and figured in passing from one skeletal region to another. For example, the humerus of *Sphenodon* is figured, but not the femur, and different aspects of the skulls of many families or suborders are represented by different genera. Some inconsistency is forced by incompleteness of the fossil remains, but it seems unfortunate that the skull of a widely known form like *Dicynodon* could not have been fully illustrated. Larger scale illustrations of the crocodilian skull are needed to adequately show the details mentioned in the text, and additional figures of dinosaur braincases, beyond figure 72, would be desirable.

To the paleontologist or herpetologist confronted with the problem of identifying reptilian skeletons, the systematic portion of the volume, with its detailed osteological diagnoses, will be most valuable. Distinctive characters are listed for every family and higher group, and for subfamilies, where these are significant. Because of their terseness, one is inevitably referred back to the section on descriptive osteology for additional data. Yet neither section of the book could have been expanded to include all that is in both parts without becoming an unwieldy mass of but dimly related details.

The lists of genera and species of reptiles, both living and extinct, together with all synonyms, would in themselves form a volume of major importance. A type of compilation particularly liable to errors, and inevitably subject to expert criticism based upon differences of opinion, this section probably is as complete and correct a statement of current reptilian taxonomy as man can produce. Its usefulness will by no means be confined to specialists in herpetology, for whom it is an essential tool.

Professor Romer's classification is conservative, and differs only in minor refinements from that outlined in his 1945 textbook "Vertebrate Paleontology". It maintains the same sense of proportion found in his widely followed classification of all vertebrates, avoiding overemphasis of the taxonomic value of minor morphological distinctions. Discussions of phylogenetic relationships explain the classification followed and amplify the treatment in "Vertebrate Paleontology" by their fuller exposition of perplexing problems. Only a few

controversial points or novel suggestions can be mentioned here: A saner taxonomy of turtles is achieved by use of superfamilies and reduction of the number of families from prevalent usage; for example the Emydidae and Testudinidae are united. *Prolacerta* is retained among the Eosuchia as not demonstrably a lizard ancestor. *Askeptosaurns* and the thalattosaurs are regarded as eosuchian off-shoots rather than primitive lizards, in contrast to Kuhn's interpretation, and gekkos are treated as specialized rather than primitive lizards.

The thorough treatment of the Protorosauria, emphasizing the distinct euryapsid features of their vertebrae and ribs as well as skull, does much to remove this order from the "dumping ground" category that it has come to be regarded, for example, in the works of Watson, Broili, and Piveteau. In contrast to E. Kuhn, Romer regards the long-necked *Tanystrophaeus* as a protorosaur rather than an early lizard. Among the mammal-like reptiles, analysis of the anomodont subdivisions appears far less critical than that of the theriodonts, but this is an area in which much basic research is yet needed.

Errors in spelling and reference citations are remarkably few for a volume of such size, and the prolonged time in press seems to be justified by the quality of the publication. In surveying the volume as a whole perhaps the greatest criticism that can be offered is that the idealized drawings, eliminating distortion and imperfections of actual specimens, while invaluable for interpretative studies, cannot replace illustrations or actual examination of individual specimens for critical analysis.

If this review has dwelt upon shortcomings of the book it is not for lack of appreciation of its merits. *Osteology of the Reptiles* is far more than its title suggests; it is a thorough survey not only of the skeleton but also of the classification and relationships of all reptiles. A year's use has proven it to be most definitely essential.

JOSEPH T. GREGORY

REVIEWS

Pollen and Spore Morphology/Plant Taxonomy; by G. ERDTMAN. Gymnospermae, Pteridophyta, Bryophyta. An Introduction Palynology II. P. iv, 151; frontispiece, 5 plates, 265 figs. New York, 1957 (Ronald Press Company, $8.00).—The standards established in Volume I of this series have been maintained in this contribution from the Palynological laboratory in Stockholm. This volume is of particular interest, in that it focuses attention on plant groups whose pollen and spore morphology are little known. It will be of value not only to pollen and spore morphologists, for whom the book was prepared, but also to investigators applying palynology to special problems.

As stated in the preface, Volume II of An Introduction to Palynology consists principally of illustrations, and the related text is to be presented in Volume III (together with new additions to Volume I). This is a rather disappointing arrangement, for it necessitates the use of both volumes simultaneously in any field in which either is needed. An impressive amount of material has been assembled in this work. The genera are arranged alphabetically by phyla. Among the Gymnospermae twelve families are represented, including fifty-seven genera. The section on Pteridophyta includes twenty-nine families and one hundred and fifteen genera. The Bryophyta are represented by twenty-

three families of Hepaticae and forty families of Musci with a total of seventy-seven genera. Photomicrographs and electron-micrographs have been used freely in conjunction with the usual palynograms. Of especial interest are the two supplementary chapters at the end of the book. The first, "On new methods in physical cell research and their application in studies of pollen grains and spores," by B. M. Afzelius is a comprehensive survey of the techniques and results of electron microscopy, polarization microscopy, X-ray diffraction, Ultra-violet microscopy, and interference microscopy. The second supplement, "On the cutting of ultra-thin sections," by J. Radwan Praglowski is a careful, step-by-step instruction in the gelatine capsule-methacrylate technique for the embedding, sectioning, and mounting of ultra-thin sections of pollen and spore material.

The few typographical errors are of little consequence. Apparently seventeen members of the Podocarpaceae were inadvertently omitted from the list of the pollen grains of the Gymnospermae. Plates IV and V face p. 130 and p. 131, respectively and not p. 108 and p. 109 as indicated in the list of illustrations for the Bryophyta. There are relatively few new terms in the book, for which investigators not familiar with the Erdtman terminology will be grateful. In this reviewer's opinion, "4-lete" or even "quadrilete" is much more manageable and just as meaningful as "tetrachotomalaesurate." Although the dimensions of the spores and pollen will undoubtedly appear in Volume III, the usefulness of the photographs and diagrams would be increased if the measurements had been included in the legends for the illustrations.

J. GORDON OGDEN, III

Encyclopédie Française, Tome III, Le Ciel et la Terre; Directors, ANDRE DANJON (for Astronomy), PIERRE PRUVOST (for Dynamical and Historical Geology), and JULES BLACHE (for Geography and Geomorphology). P. 448; 35 pls., many text figures. Paris, 1956 (Société de l'Encyclopédie française; Larousse).—This ambitious volume has set for its object the explanation of the sciences of the Earth and the Heavens to the cultured layman, an explanation written not by journalists or popularizers but by professional scientists. In all, 36 names are listed in the index of collaborators, all Frenchmen of recognized standing in their science. No compromise seems to have been accepted that would sacrifice either accuracy or depth, save possibly what was absolutely necessary to compress so vast a subject into a quarto volume of less than 450 pages.

Part I, The Heavens (144 p. and 24 pls.), is divided into four sections: The Earth and the Solar System, The Sun and the Stars, The Galaxies, and Problems of Evolution. Part II, The Earth, is divided into three: The Terrestrial Globe (52 p.), The Crust of the Earth (88 p., 2 pls.), and The Economy of the Earth's Surface (138 p., 9 pls.); of these the first is essentially geophysics, the second dynamical and historical geology, the third physical geography in the broadest sense, i.e. climatology, physical oceanography and limnology, etc., and geomorphology.

To give an idea of the scope of the volume, let us concentrate on one section only: The Crust of the Earth, written by members of the faculty of geology at the Sorbonne. It begins with a brief introduction by Prof. Pierre

Pruvost, discussing particularly the historical rise of geology, with appendix on absolute age measurement. The first chapter, by Prof. Louis Glangeaud, is a relatively short (12 p.) but exceedingly concentrated discussion of the place of geology among the sciences, with special emphasis on the different levels of organization of matter on different scales from the molecule to the galaxy. The second chapter, by Dr. Pierre Routhier, forms more than half of the whole section; entitled The Formation of the Materials of the Crust, it is an extended essay on petrology, emphasizing principles and ideas almost to the exclusion of conventional rock description and classification. In a final chapter, Prof. Pruvost covers historical geology in 14 pages, plus 2 plates of paleogeographic maps of Europe. (A chapter on deformations of the crust was also planned but not completed; the loose-leaf plan of the encyclopedia will permit its later insertion.)

For the professional geologist, these pages provide an unexcelled opportunity to become familiar with the basic ideas now current in French geology, as seen by teachers deeply interested in the interrelations of those ideas. In each case, indeed, a new synthesis of the ideas is in effect attempted. As for the layman, however, one wonders whether the concentrated style and the constant allusion to geological terms and concepts not at once explained will not baffle him long before he has completed a section, and whether, if he does persevere, the rather individual character of the syntheses will prepare him well for further reading in geology.

Nevertheless one cannot help admiring these syntheses and the enthusiasm, not to say audacity, that went into preparing them for the cultured layman, to make him see the philosophical meaning and interconnectedness of geology, so easily lost sight of in its increasing specialization. If France possesses a large audience for such scientific writing, we also ought to reexamine our usual assumptions about what the layman can understand and try to determine whether such an audience may not exist here.

JOHN RODGERS

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